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Stratigraphy and Sedimentology of the Upper Silurian strata on eastern Prince of Wales Island, Arctic Canada

by

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Stratigraphy and Sedimentology of the Upper Silurian strata on eastern Prince of Wales Island, Arctic Canada submitted by Paul Stuart Mortensen in partial fulfilment of the requirements for the degree of Doctor of Philosophy in Geology.

DEDICATION

Til resten av trolla under brua.

Abstract

The Upper Silurian stratigraphic sequence on eastern Prince of Wales Island represents a transgressive-regressive cycle with the alluvial deposits of the overlying Peel Sound Formation marking the upper limit of the regression. Three formations are encompassed by this study; the Cape Storm, the Douro and the Somerset Island formations.

The strata of the Cape Storm and Somerset Island formations represent deposition on carbonate-dominated tidal flats. Variably quartzose dolostones dominate the supratidal-intertidal zones of these tidal flats. The quartz grains and some of the dolomite grains are terrigenous and derived from the Boothia Horst to the east. Planar-bedded, mottled dolomitic biomicrites mark the transition to and from the continuous sequence of rubbly weathering, mottled dolomitic limestones of the Douro Formation.

The Douro Formation contains a diverse open marine fauna and represents the maximum transgression during the Ludlovian. During deposition of the Douro Formation, the Boothia Horst was a shallow submergent feature separating the M'Clintock basin to the west and the Prince Regent Basin to the east. The strata of the Douro Formation are unique relative to the other strata of this study in that they lack significant amounts of terrigenous detritus.

Mottled dolomitic limestones (biomicrites and biopelmicrites) comprise approximately half of the strata. Three types of mottled dolomitic limestones are recognized

on the basis of the continuity of the micrite lumps or layers relative to the distribution of the idiotopic dolomitic matrix. The mottled dolomitic biomicrites are largely biogenetic in origin and form a progression of genetically related rock types. The resultant rock type is dependent on the relative rates of sedimentation, bioturbation and early lithification.

The strata contains an abundant algal flora of Schizophyta, Rhodophycophyta and Chlorophycophyta. Hemispherical and undulatory stromatolites are the dominant growth forms in the Cape Storm Formation and were common only on the transgressive tidal flats. Oncoliths comprised of *Girvanella* were abundant during the regressive phase of the lower part of the Somerset Island Formation. Rhodoliths (nodules of the red algae *Solenopora*) are abundant and show a similar stratigraphic distribution as the mottled dolomitic limestones. Morphological differences between and within the two spherical growth forms indicate abiotic factors controlled morphology.

The fauna is dominated by the brachiopod genus *Atrypoidea* of which four species are recognized. These species form a well-defined zonation that occurs throughout the Arctic Lowlands.

The Boothia Horst had a pronounced effect on the geological history of the area, both as a sediment source and as a control on sedimentation patterns. Substantial thickness and facies variations in the stratigraphy, in

particular the mottled dolomitic limestones of the Douro Formation, are attributed to the differential subsidence of fault-bounded sub-basins of the M'Clintock Basin. Uplift during the Early Devonian resulted in the Cornwallis Fold Belt; the structural province in which the strata of this study occur.

ACKNOWLEDGEMENTS

This thesis is the product of the interactions with all the people I have encountered over the years and to whom I express my gratitude.

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Moral support of this thesis was generously given by my family and friends. In particular, I would like to express my thanks to three women; my wife Phil, my Mom and Helen Lee for their love and support throughout these years. A special thanks to Phil for her understanding during the dark and surly days of the last year.

I would like to thank Brian Jones who served as my academic advisor, counsel and friend for his time, patience and help during the gestation of this thesis. Hilary Vaughan and Jill Peacocke contributed their services as my field companions, offering their friendship and tolerance under sometimes adverse conditions. Drs. B. Chatterton and T. Uyeno kindly identified the conodonts.

My sincerest thanks to Dr. C.R. Stelck for sharing his wisdom and humor with me. Much of my understanding of the mysteries of geology was acquired in the realm of Rm. 193. Finally, I would like to thank my examining committee of Drs. G. Pemberton, C. Kahle, S. Thomson, C. Stelck and B.

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I. INTRODUCTION

A. General statement

Innumerable studies of tidal flat and shallow water carbonates are documented in the literature. In more accessible (and more hospitable) areas, detailed geological investigations have been ongoing for decades and are often into their second century. This, however, is not the situation in the Canadian Arctic Archipelago. Geological observations of this area were made by the early explorers but it was not until the mid-twentieth century that expeditions solely designated for geological investigations were initiated. Since that time, much of the geologic history has been elucidated; however, voids in our knowledge still exist.

The intent of this study was to examine the Upper Silurian strata on eastern Prince of Wales Island. Located in the M'Clintock Basin, it is the least studied basin of the Arctic Lowlands. While the objectives of this study were defined prior to its initiation, it was the nature of the rocks that ultimately dictated the direction this study followed. The proximity of the basin margin to the Boothia Horst was evident from the initiation of this study but the pronounced influence of the Boothia Horst on the Ludlovian geological history could not be anticipated. Neither could the abundant algal flora nor the importance of biogenetic (organosedimentary) structures be anticipated. These

features are relatively undocumented in the Upper Silurian strata of the Arctic Lowlands. The following text is a documentation and interpretation of these features and their contribution to the geological history of the Arctic Lowlands.

B. Scope of the present study

The area of study is eastern Prince of Wales, Pandora, Prescott and Vivian islands between 72°07'N latitude and 73°16'N latitude (Fig. 1.1). Initially, the study extended north to Cape Hardy (73°54'N latitude) but due to very poor exposure north of Browne Bay the study was restricted to the southern area as specified above. Exposures of the strata are restricted to a narrow belt along eastern Prince of Wales Island and the adjacent islands between 96°24'W and 96°55'W longitude. The vertical to overturned strata delineate the western margin of the Cornwallis Fold Belt.

The primary objectives of the present study are:

1. Delineation of the stratigraphy on eastern Prince of Wales Island and the adjacent islands of Pandora, Prescott and Vivian.
2. Delineation of the facies in the Upper Silurian strata of the study area.
3. Production of a detailed geologic map at a scale of 1:50,000 of the study area.
4. Establish and describe the Upper Silurian geological history and paleogeography of the study area.

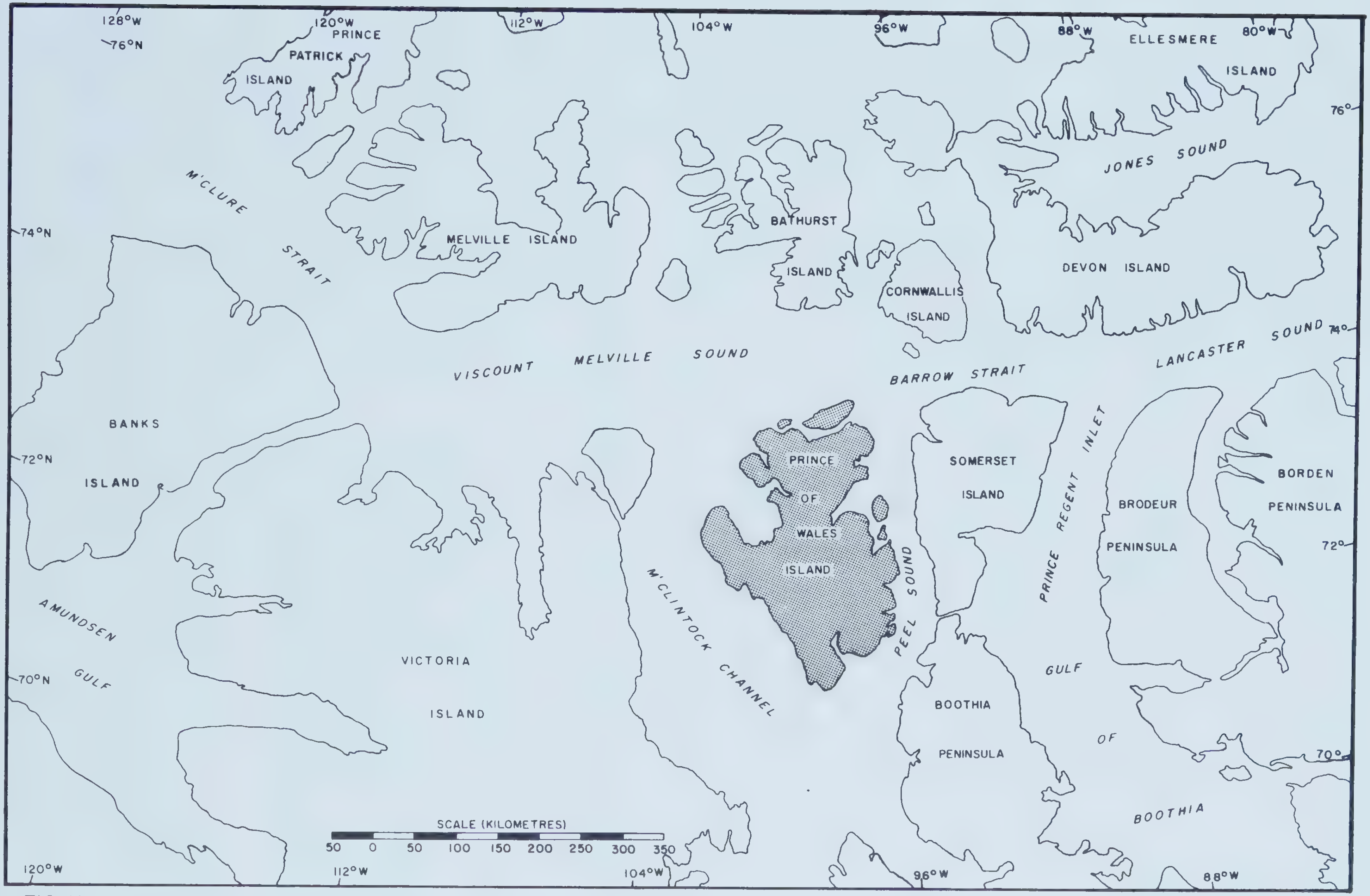


FIG. I.I. Location of Prince of Wales Island in the Canadian Arctic Archipelago.

5. Examine the role of the Boothia Horst on sedimentation in the M'Clintock basin during the Ludlovian.

C. Location and access

Prince of Wales Island is located in the Canadian High Arctic between latitudes 71° and 74°N and longitudes 96° and 103°W (Fig. 1.1). Topographic relief is greatest along the eastern margin of Prince of Wales, Pandora and Prescott islands; a maximum elevation of 330 metres occurs over the Precambrian terrain exposed along the eastern coast of the forementioned islands. The relief over most of Prince of Wales Island is less than 60 metres. Unless otherwise specified, further reference to Prince of Wales Island will be assumed to include Pandora, Prescott and Vivian islands (Fig. 1.2).

These islands are uninhabited and the nearest permanent settlement is Resolute Bay on Cornwallis Island. Transport from Resolute Bay to Prince of Wales Island was by twin-engine Otter aircraft chartered by the government-sponsored Polar Continental Shelf Project. Travel on the island was by foot or by limited use of a Bell 206B helicopter. Both fixed wing and rotary aircraft were used to relocate the camps on the island.

Standard vertical air-photos at a scale of approximately 1:50,000 were used for location of camps and geologic mapping.

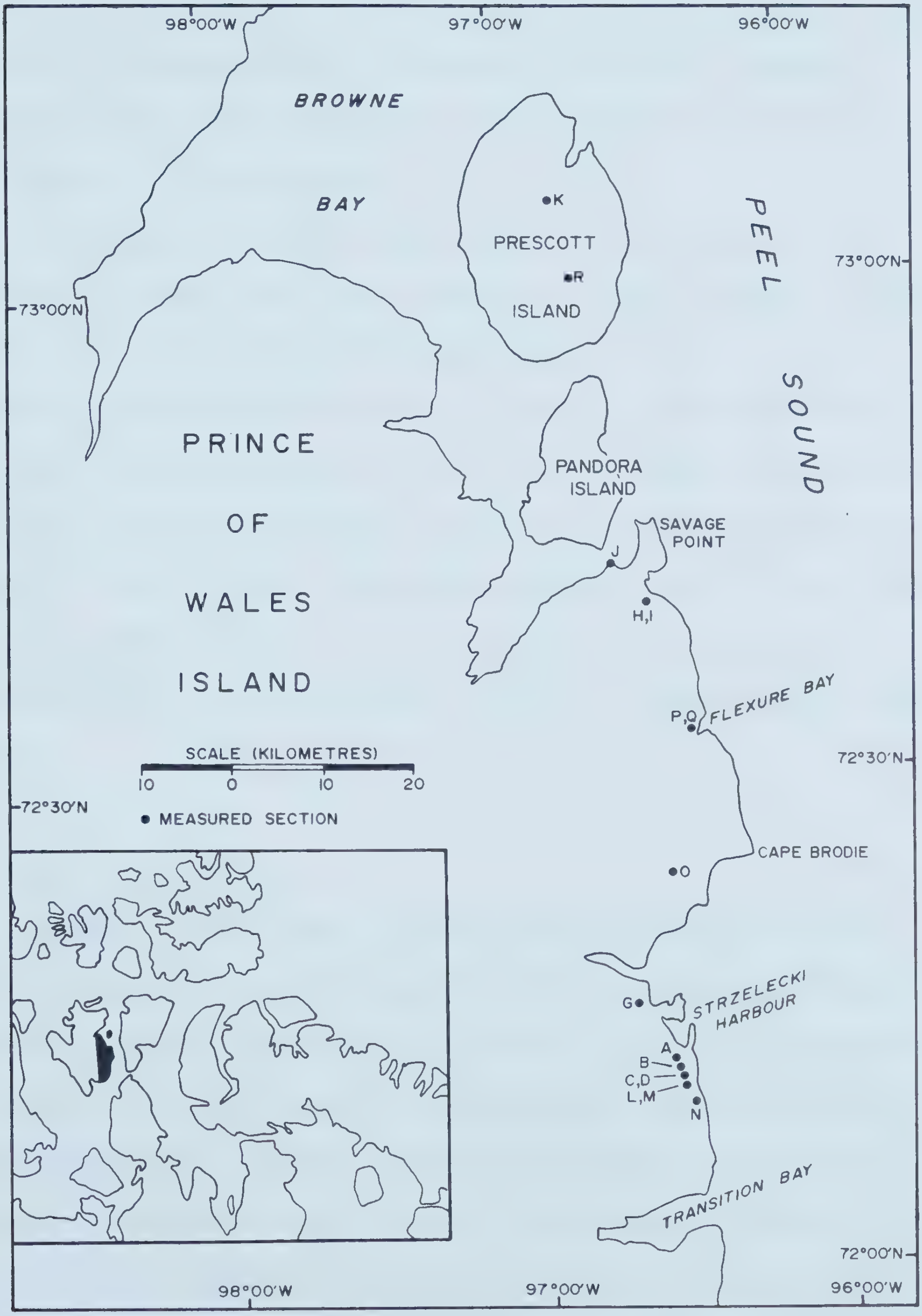


FIG. 1.2. Location of measured sections on Prince of Wales and Prescott Islands.

Most of the land surface is covered by glacial drift or felsenmeer. The felsenmeer is confined to a narrow strip of overturned strata of the Cornwallis Fold Belt along eastern Prince of Wales, Wales, Prescott and Pandora islands. Glacial drift obscures much of the bedrock geology over most of Prince of Wales Island.

Measurement of geologic sections was along river gorges, coastal sections or areas of good surface exposure (Fig. 1.2). Mapping was accomplished by the use of air photos, measured stratigraphic sections and traverses of up to a 40 kilometre radius from the base camps. The felsenmeer is essentially weathering *in situ* and could, therefore, be used for mapping the formations. During the 1982 and 1983 field seasons, Prince of Wales Island and the adjacent islands were traversed along the full length of the study area.

D. Previous work

Prior to 1955 when the Geological Survey of Canada staged a major reconnaissance project (Operation Franklin; Fortier, *et al.*, 1963), no systematic study of the geology of Prince of Wales Island and adjacent islands had been done. Operation Franklin extended only as far south as Pandora Island. During 1960 and 1961, Dominion Explorers Ltd. examined the lower Paleozoic strata on eastern Prince of Wales Island. A total of 12,000 metres (40,000 feet) of lower Paleozoic strata were measured and a brief summary

report was released through the Department of Indian and Northern Affairs (McNair, 1962).

On the basis of airborne magnetic and radiometric profiles flown over Prince of Wales Island, Gregory *et al.* (1961) estimated the depth to the Precambrian basement to be between 1500-3000 metres (5000-10,000 feet) in the southeastern part of the island and probably greater than 3000 metres (10,000 feet) beneath the rest of the island. This thickness may include Proterozoic strata in addition to Paleozoic strata. Also resulting from this survey was the recognition of a regional basement fault along the western margin of the Boothia Arch on eastern Prince of Wales Island and in the Peel Sound (Gregory *et al.*, 1961).

In 1962, The Geological Survey of Canada staged a second major reconnaissance project, Operation Prince of Wales. This study encompassed Prince of Wales, Somerset, and King Williams islands plus the Boothia Peninsula. This study resulted in a preliminary report and the first geological map of this area (Blackadar and Christie, 1963).

The University of Ottawa had established a series of research programs on Prince of Wales Island. Broad (1969) examined the ostracoderms from the lower member of the Peel Sound Formation and Miall (1969) studied the Peel Sound Formation. The Cambrian to Lower Silurian strata was examined by Dixon (1973a) and Williams (Jones, personal communication, 1981) initiated a study on the Read Bay Formation.

In 1975, Kerr McGee Ltd. drilled a well (KMG Decalta Young Bay F-62) on Prince of Wales Island at 72°41'23''N and 96°49'34''W. The oldest strata penetrated by this well was the Proterozoic Aston Formation. Other exploration interest has been confined to seismic exploration (Released Geophysical and Geological Reports-Canada Lands, Canadian Oil and Gas Lands Adminstration, 1983). Since 1975, Prince of Wales Island has not been the subject of any detailed geological investigation until this study was initiated in 1981.

Three field seasons (1981, 1982, 1983) were spent on eastern Prince of Wales Island and nine camps were located. A total of twelve sections were measured (Appendix I) and the locations of the camps are approximately coincident with the location of measured sections. The length of the field seasons ranged from 5-8 weeks in duration; a time span corresponding to the length of the brief summer in the high latitudes.

II. REGIONAL GEOLOGY

A. Structural provinces

Rocks exposed on eastern Prince of Wales Island range from Archean crystalline rocks through to and including Early Devonian clastic sedimentary rocks (Figs. 2.1 and 2.2). Based on the geological history of the area and the present-day outcrop pattern, Prince of Wales Island can be subdivided into three structural provinces namely;

The Boothia Horst: A north-south trending belt of Proterozoic (Aphebian) metamorphic rocks (Fig. 2.3). Only small remnants of the Boothia Horst are exposed along the eastern coastal margin of Prince of Wales, Prescott, and Pandora islands (Figs. 2.1 and 2.2).

The Cornwallis Fold Belt: This is a band of overturned strata along the margins of the Boothia Horst (Fig. 2.3). This belt is less than 3 kilometres wide on eastern Prince of Wales Island and resulted from tectonic pulses associated with the Boothia Horst during the lower Paleozoic (Kerr and Christie, 1965; Kerr, 1977). The strata encompassed by the present study occur in the western margin of the Cornwallis Fold Belt. The general structure of the western margin of the Cornwallis Fold Belt on eastern Prince of Wales Island is that of a single asymmetrical syncline with subsequent erosion of most of the overturned limb. The Cornwallis Fold Belt and the Boothia Horst comprise the Boothia Uplift.

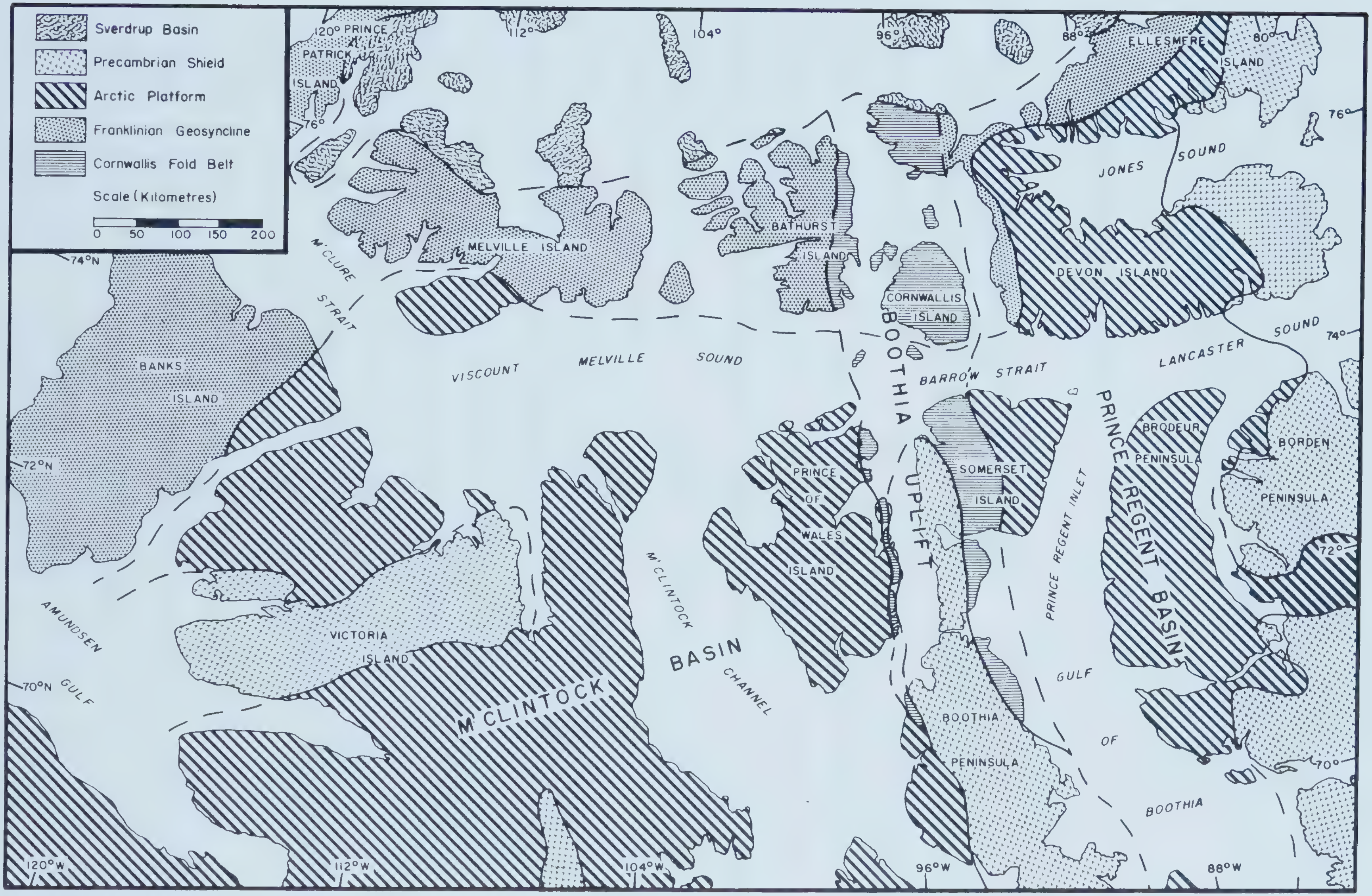


FIG. 2.3. Major structural provinces of the central Canadian Arctic Archipelago (from Tozer and Thorsteinsson, 1963; Miall and Kerr, 1980 and Thorsteinsson, 1980).

The M'Clintock Basin: Essentially undeformed strata of the Arctic Lowlands (Fig. 2.3). Most of Prince of Wales Island lies in the M'Clintock Basin which flanks the western margin of the Cornwallis Fold Belt.

B. Regional Stratigraphy

Lang River Formation and older units

The basement of the study area is comprised dominantly of Aphebian metamorphic rocks, predominantly mafic schists and felsic gneisses, plus small granite and ultrabasic bodies (Blackadar, *in* Blackadar and Christie, 1963). The outcrop belt of the basement extends along a narrow coastal strip of eastern Prince of Wales, Prescott and Pandora islands (Figs. 2.1 and 2.2).

The Aphebian granitic gneisses are unconformably overlain by Proterozoic sedimentary rocks (Chandler, 1980, p. 27). The lowest stratigraphic occurrence of sedimentary rocks are the silica cemented red quartzites, red shales and stromatolitic dolosiltites of the Aston Formation (Table 2.1) (Tuke *et al.*, 1966). Maximum thickness of the Aston Formation is 800 metres (Chandler, 1980).

Conformably overlying the Aston Formation is the Hunting Formation. The maximum thickness of the Hunting Formation is 2250 metres (Chandler, 1980). Tuke *et al.* (1966) recognized four members in the Hunting Formation: 1) a basal dark brown shale and orange dolostone member, 2) a

AGE	FORMATION	THICKNESS	LITHOLOGY
Upper Silurian-Lower Devonian	Peel Sound Formation		
	upper member	300+ m	conglomerate, conglomeratic sandstone, limestone and dolostone.
	lower member	120-200 m	sandstone and conglomerate.
Upper Silurian	Somerset Island Formation	115-257 m	variably quartzose dolomitic limestone, dolostone, quartz arenite and siltstone.
Upper Silurian	Douro Formation	0-262 m	mottled dolomitic limestone.
Upper Silurian	Cape Storm Formation	0-97 m	variably quartzose dolostone and dolomitic limestone.
Upper Ordovician-Middle Silurian	Allen Bay Formation	0-741 m	dolostone.
Middle Cambrian-Upper Ordovician	Lang River Formation	0-214 m	dolostone and sandstone.
	?unconformity?/?fault?		
Proterozoic	Hunting Formation	2250 m	dolostone.
	Aston Formation	800 m	quartz arenite, dolostone and shale.
	unconformity		
Aphebian	Aphebian crystalline rocks		mafic schists, felsic gneisses, granite and gabbro.

TABLE 2.1. Stratigraphy on eastern Prince of Wales Island (from Tuke *et al.*, 1966; Miall, 1969; Dixon, 1973; Chandler, 1980; this study).

grey dololutite, dolosiltite and chert member, 3) a stromatolitic dolostone member and 4) a marly dolostone and stromatolitic dolostone member. Both the Aston and Hunting formations are intruded by gabbroic sills and dykes.

The age of the unfossiliferous Aston and Hunting formations is open to question. Initially, these formations were assigned a Middle Cambrian to Ordovician age based on lithologic similarity with known Paleozoic rocks in the northern Boothia Peninsula and the presence of an unconformity with the overlying Ordovician strata (Tuke *et al.*, 1966). Assignment of a Proterozoic age is based on similarity in lithology and succession to other presumed Proterozoic sequences in the Arctic (Tuke *et al.*, 1966). Similar sandstones are found on Ellesmere Island and Greenland and are overlain by Cambrian fossil-bearing strata (Tuke *et al.*, 1966).

The Lang River Formation (Dixon, 1973b) on Prince of Wales Island is characterized by two types of cyclothems: dolostone-sandstone cyclothems at Savage Point and fissile dolostone-stromatolitic dolostone cyclothems at Strzelecki Harbour. On Somerset Island, the Lang River Formation unconformably overlies Proterozoic strata or metamorphic basement whereas on Prince of Wales Island the Lang River Formation and younger formations are faulted against Precambrian rocks (Dixon, 1973b, p. 134). No complete section of the Lang River Formation occurs on Prince of Wales Island and a maximum measured thickness of the

formation is 214 m (Dixon, 1973b). The Lang River Formation is unfossiliferous on Prince of Wales Island and is assigned a Middle Cambrian to Upper Ordovician age (Dixon, 1973a, 1973b).

Allen Bay Formation

The Allen Bay Formation was named for a succession of on Cornwallis Island (Thorsteinsson and Fortier, 1954). Later, Thorsteinsson (1958) defined the formation as a uniform succession of dominantly dolomitic rocks which overlies the Cornwallis Formation with a sharp but conformable contact and is gradationally overlain by the Read Bay Formation. The type section is along an unnamed creek (75°45'55"N, 95°05'30"W) that flows into Allen Bay on western Cornwallis Island.

The Allen Bay Formation is a sequence of thin to thick bedded, flaggy to massive, fine to coarsely crystalline dolostones (Thorsteinsson and Fortier, 1954). Dolostone comprises greater than 90% of the formation with dolomitic limestone and limestone being the remaining constituents. The formation typically weathers yellowish to brownish-grey. Poorly preserved corals and stromatoporoids dominate the sparse fauna. The Allen Bay Formation is Upper Ordovician to Middle Silurian in age (Thorsteinsson and Fortier, 1954; Thorsteinsson, 1958; Dixon, 1973a; 1973b).

Christie, (*in* Blackadar and Christie, 1963) noted a sequence of dolostones, dolomitic sandstones, and sandstones

on eastern Prince of Wales Island which are equivalent to the Cornwallis and Allen Bay formations on Cornwallis Island. Dixon (1973a, 1973b) recognized the Allen Bay Formation on Prince of Wales Island and redefined the Allen Bay Formation as a buff dolostone, generally thick bedded to massive and resistant to weathering. He placed the upper formation boundary at the stratigraphically highest resistant thick bedded to massive dolostone. Above this, Dixon (1973a, 1973b) recognized a transitional sequence of recessive, thin bedded, planar and cross-laminated dolostones and limestones which he assigned to the Read Bay Formation. These units are now assigned to the Cape Storm Formation (Kerr, 1975). Thorsteinsson (1980) amended his original description and redefined the Allen Bay Formation as a thick bedded to massive, medium to coarsely crystalline, resistant dolostone.

In this study, the upper boundary of the Allen Bay Formation as defined by Dixon (1973a,1973b) and Thorsteinsson (1980) is recognized as the Allen Bay Formation-Cape Storm Formation contact. Where possible it is used as the datum from which stratigraphic sections were measured.

Cape Storm Formation

Kerr (1975) named the Cape Storm Formation for a succession of limestones and dolostones that had previously been included in either the Allen Bay Formation or in the

Read Bay (=Douro) Formation. On Prince of Wales Island, this strata was originally assigned to the Allen Bay Formation (Christie *in* Blackadar and Christie, 1963). Dixon (1973a, 1973b), however, assigned this strata to the Read Bay Formation.

The type section of the Cape Storm Formation is on southwestern Ellesmere Island where Kerr (1975) recognized two members, a lower resistant cliff forming limestone and an upper thin bedded dolostone to silty dolostone. In the type section, the upper member is evident where the rocks abruptly become thin to medium bedded, sandy and silty, recessive dolostones that weather yellowish (Kerr, 1975). On Devon Island, Morrow and Kerr (1977) identified three lithofacies, a limestone interclast facies, a stromatolite facies, and a dolomite pelmicrite facies. On Somerset Island, Miall and Kerr (1977) recognized a sequence of laminated dolostones, sandy dolostones, and limestones with interclast breccias, ripple marks, stromatolites and a limited fauna of gastropods, ostracodes, and eurypterids which they assigned to the Cape Storm Formation.

In this study, three informal members of the Cape Storm Formation are recognized on eastern Prince of Wales Island: a lower dolostone/quartzose dolostone member, a stromatolitic dolostone/algal laminated dolostone member and an upper dolostone and dolomitic limestone member. The members are variable in thickness and lateral continuity throughout the outcrop belt. The Cape Storm Formation on

eastern Prince of Wales Island has a maximum thickness of 97 metres (Table 2.1).

On eastern Prince of Wales Island, the base of the Cape Storm Formation is placed at the lowest stratigraphic occurrence of a thin to medium bedded dolostone. In areas of poor exposure, this boundary is marked by a distinct break in slope between the resistant Allen Bay Formation and the overlying recessive Cape Storm Formation (Thorsteinsson, 1980). Gradationally overlying the Cape Storm Formation is the Douro Formation.

The Cape Storm Formation is Llandoverian to Ludlovian in age on Ellesmere Island (Kerr, 1975) and is early to late Ludlovian in age elsewhere (Thorsteinsson, 1980). On eastern Prince of Wales Island, the Cape Storm Formation of middle to late Ludlovian in age (Thorsteinsson, 1980).

Douro Formation

Thorsteinsson (1963) proposed the Douro Formation for a sequence of mainly grey limestones exposed on northwestern Devon Island. The sequence lies conformably between the rocks then regarded as the Allen Bay Formation and the Devon Island Formation (Thorsteinsson, 1963). The type section is in the Douro Range on the northeast side of Prince Alfred Bay, Devon Island. Thorsteinsson (1963, p. 227) noted the similarities in age, lithologic, and faunal characteristics of member A of the Read Bay Formation and the Douro Formation.

Thorsteinsson and Fortier (1954) recognized four members (members A, B, C, and D) in the Read Bay Formation on Cornwallis Island. Later, Thorsteinsson (1958, p. 48) suggested that these four members of the Read Bay Formation could be elevated to formational status. Thorsteinsson and Tozer (1962, p. 45) made reference to the Read Bay Group but it was not until 1980 that Thorsteinsson formally elevated the formation to group status. The Read Bay Group comprises three formations; the lower Douro Formation which equates with member A, the overlying Barlow Inlet Formation which equates with members B and C, and the Sophia Lake Formation which equates with member D (Thorsteinsson, 1980).

Christie (*in* Blackadar and Christie, 1963) recognized an undifferentiated Read Bay Formation on eastern Prince of Wales Island. The lower transitional sequence assigned to the Read Bay Formation by Dixon (1973a, 1973b) was placed in the Cape Storm Formation (Thorsteinsson, 1980). Thorsteinsson (1980, p. 4) equated, both lithologically and chronologically, all of the Read Bay Formation on Somerset Island (Miall and Kerr, 1977; Jones *et al.*, 1979) and on Prince of Wales Island with the Douro Formation. The revised stratigraphy on eastern Prince of Wales Island (this study) assigns some of the upper beds of the Read Bay Formation (Miall, 1969) and the lower beds of the Peel Sound Formation (Thorsteinsson, 1980) to the Somerset Island Formation.

On eastern Prince of Wales Island, a break in slope commonly marks the contact between the Cape Storm and Douro

Formations. The thin bedded dolostones of the Cape Storm Formation generally weather recessive while the Douro Formation is more resistant. The base of the formation is placed at the lowest stratigraphic occurrence of a continuous sequence of rubbly weathering limestones (Thorsteinsson, 1963; 1980). The Douro Formation on eastern Prince of Wales Island has a maximum thickness of 262 metres (Table 2.1).

Jones and Dixon (1977) and Jones *et al*. (1979) recognized two dominant lithologies in the Read Bay Formation (=Douro Formation) on Somerset Island; rubbly argillaceous limestones and rubbly dolomitic limestones. Only the latter is dominant on eastern Prince of Wales Island. The Douro Formation is fossiliferous containing brachiopods, corals, stromatoporoids, trilobites, coralline algae, gastropods, ostracods, echinoderms and nautiloids.

The Douro Formation is late Ludlovian in age throughout the Arctic Archipelago (Thorsteinsson, 1980). On Somerset Island, Jones and Dixon (1977) assigned a late Ludlovian to middle Pridolian age to the Read Bay Formation whereas on eastern Prince of Wales Island, the Douro Formation is late Ludlovian in age (Fig. 2.4). On eastern Somerset Island, Jones and Dixon (1977) and Miall and Kerr (1980) regarded the Read Bay Formation as being a diachronous unit whereas Thorsteinsson (1980) regarded the Douro Formation as being a synchronous unit throughout the Arctic Archipelago. Conodont faunas, collected during this study, are from the late

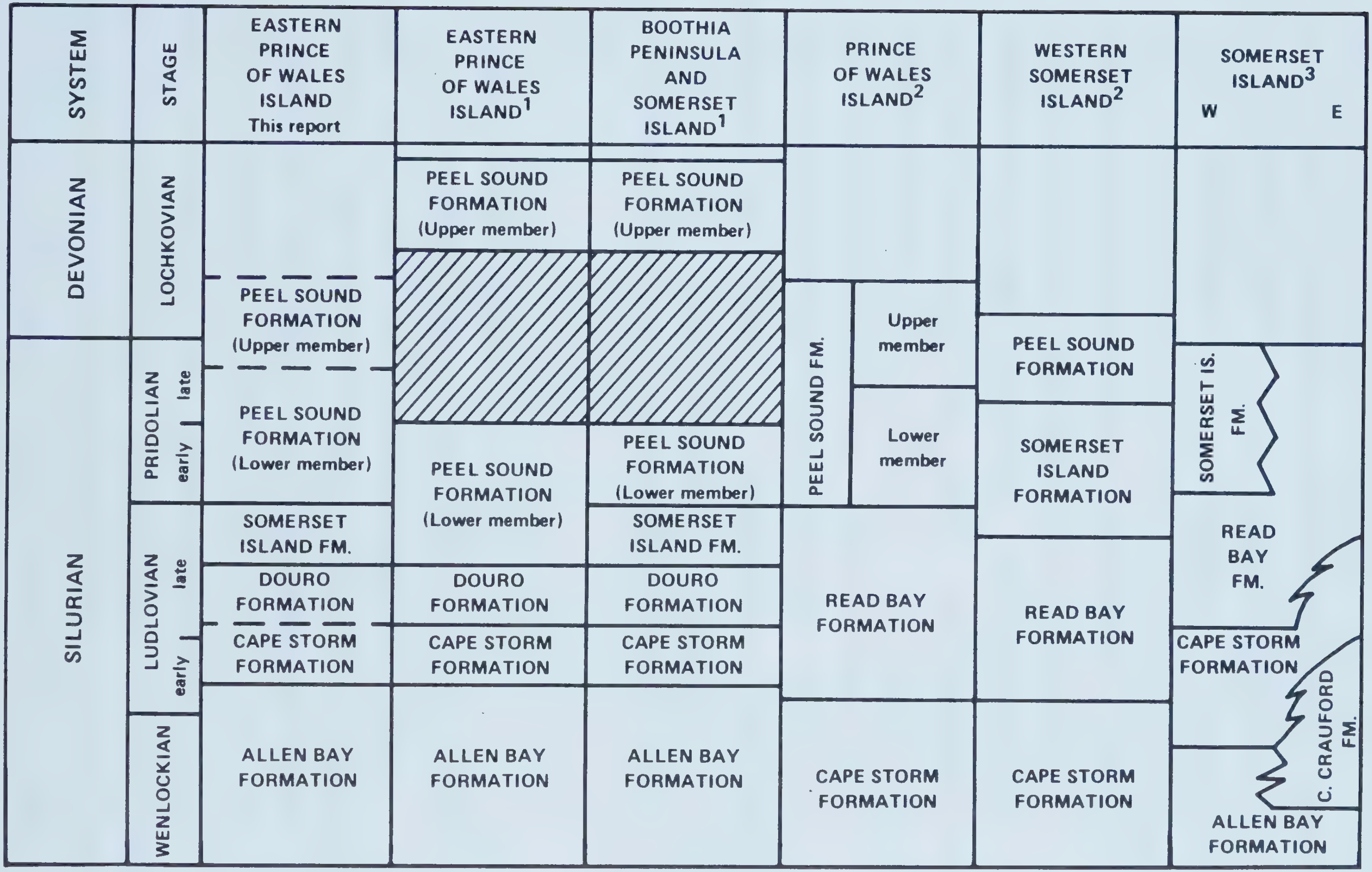


FIG. 2.4. Regional stratigraphy of the Upper Silurian in the south-central Arctic Archipelago (from 1. Thorsteinsson, 1980; 2. Miall *et al.*, 1978; 3. Miall and Kerr, 1980).

Ludlovian *siluricus* and *latialata* zones (Appendix II). The Douro Formation is possibly diachronous along the outcrop belt on eastern Prince of Wales Island (Appendix II).

The Douro Formation is gradationally overlain by the Somerset Island Formation. At Bellot Cliff on northeastern Prince of Wales Island, there is an angular unconformity between the Read Bay Formation (=Douro Formation of Thorsteinsson, 1980) and the Peel Sound Formation (Kerr and Christie, 1965; Miall, 1969). Miall (1969) assigned the conglomerate above the unconformity to the lower Peel Sound Formation whereas Thorsteinsson (1980) assigned this conglomerate to the base of the upper Peel Sound Formation. The author did not visit this locality and it is unknown as to what portion of the stratigraphic succession is missing with regards to the revised nomenclature of this study (Fig. 2.4).

Somerset Island Formation

The Somerset Island Formation (Miall *et al.*, 1978) was named for a transitional sequence between the rubbly weathering limestones and dolostones of the underlying Douro Formation and the conglomerates and sandstones of the overlying Peel Sound Formation on Somerset Island. The distinctive nature of this transitional sequence was recognized by many workers (Thorsteinsson and Tozer, 1963; Christie, *in* Blackadar and Christie, 1963; Dineley, 1966; Brown *et al.*, 1969; Miall, 1970a; Reinson *et al.*, 1976;

Gibling and Narbonne, 1977; Jones and Dixon, 1977; Miall and Kerr, 1977). This sequence has previously been assigned to either the upper Read Bay Formation and/or the lower Peel Sound Formation (Thorsteinsson, 1980, p. 9).

On Somerset Island, the Somerset Island Formation equates with the unnamed Upper Silurian sequence of Reinson *et al.* (1976), Gibling and Narbonne (1977), Jones and Dixon (1977), and Miall and Kerr (1977). This transitional sequence correlates with member D of the Read Bay Formation on Cornwallis Island (Dineley, 1966; Gibling and Narbonne, 1977; Miall *et al.*, 1978).

The type section (76°21'N, 92°51'W) is along a tributary of the Creswell River on Somerset Island. Miall *et al.* (1978) divided the formation into two members, a lower carbonate member and an upper carbonate-clastic member. The lower member is comprised of grey, fine grained, planar-bedded and mottled limestones and dolostones. The base of the formation is placed at the lowest stratigraphic occurrence of a planar-stratified dolostone or limestone (Miall *et al.*, 1978). Thorsteinsson (1980) noted a similar boundary and placed the base of the formation at the change from rubbly weathering limestones to dominantly planar-bedded, variably quartzose, silty and sandy dolomitic rocks.

The upper member of the Somerset Island Formation is comprised of dolosiltite, quartzose siltstone and mudstone, and lithologies of the lower member (Miall *et al.*, 1978).

The contact between the lower and upper member is placed at the lowest stratigraphic occurrence of a red siltstone (Miall *et al.*, 1978). This coincides with the contact between the Read Bay Formation and the Peel Sound Formation as originally defined by Thorsteinsson and Tozer (1963). The top of the Somerset Island Formation on Somerset Island is placed at the lowest stratigraphic occurrence of a red-colored, cross-bedded sandstone (Miall and Kerr, 1977; Miall *et al.*, 1978). On Somerset Island, the Somerset Island Formation varies in thickness from 150 to greater than 400 metres (Miall *et al.*, 1978).

The Somerset Island Formation on eastern Prince of Wales Island (as defined herein, Fig. 2.4) is named for a succession of planar-bedded, variably quartzose limestones and dolostones with lesser amounts of quartz arenite and siltstone. The sequence is underlain by the rubbly weathering limestones of the Douro Formation and overlain by the coarse clastics of the Peel Sound Formation (Fig. 2.4). Typically, the sequence is more calcareous at the base with dolomite increasing up-section.

The base of the formation is placed at the lowest stratigraphic occurrence of a quartz arenite or a quartzose limestone or dolostone. Coincident with this contact is the change from predominantly rubbly weathering limestones to planar-bedded limestones and dolostones. This contact is commonly marked by a 12-15 metre recessive gully which can be mapped along strike in some localities. The contact

between the Douro and Somerset Island formations is sharp and conformable.

The Somerset Island Formation on eastern Prince of Wales Island incoporates strata previously assigned to both the Read Bay Formation and the lower member of the Peel Sound Formation (Thorsteinsson and Tozer, 1963; Miall, 1969; Thorsteinsson, 1980). Thorsteinsson (1980) placed the base of the lower member of the Peel Sound Formation on eastern Prince of Wales Island at the lowest stratigraphic occurrence of a quartzose sandstone. Thorsteinsson (1980) also divided the lower member of the Peel Sound Formation into two informal parts, a lower part of predominantly sandstones with lesser amounts of limestone and dolostone and an upper part of red weathering conglomerate, conglomeratic sandstone, and sandstone. The lower part of the lower member of the Peel Sound Formation (approximately 200 metres) equates with the Somerset Island Formation as defined herein. The upper part of the lower member remains in the Peel Sound Formation.

Miall *et al.* (1978) suggested that the lower 50 metres of the Peel Sound Formation and the sandstone beds in the Read Bay Formation (Miall, 1969) may be more appropriately included in the Somerset Island Formation. They also noted a 275 metre sandstone-limestone-shale succession above the Read Bay Formation in the Young Bay F-62 well. "The lower part of this succession presumably correlates with the Somerset Island Formation" (Miall *et al.*, 1978, p. 185).

The Somerset Island Formation on both Somerset and eastern Prince of Wales islands is similar. On both islands, the lower strata of the Somerset Island Formation are predominantly limestones becoming predominantly dolostones approximately half way up the formation. Mottled dolomitic limestones, similar to those in Douro Formation, are found at the base of the formation and decrease up-section. The Somerset Island Formation on eastern Prince of Wales Island contains few pronounced red weathering beds but there is a tendency for the pink-green weathering quartzose dolosiltites to increase up-section.

Although the Somerset Island Formation is similar on both Prince of Wales and Somerset islands, the two members described by Miall and Gibling (1978) and Miall *et al*. (1978) on Somerset Island are not recognized on eastern Prince of Wales Island. However, the Somerset Island Formation is subdivided into a lower and upper part that are lithologically similar to the two members. The contact between the two parts is gradational and the red siltstone separating the two members on Somerset Island (Miall *et al*., 1978) was not found on eastern Prince of Wales Island.

On both islands, the Somerset Island Formation is variably quartzose whereas the underlying Douro Formation contains virtually no coarse clastic detritus. In only one section (Section PQ) is there any significant clastic content in the Douro Formation. At approximately the middle of the Douro Formation (Q-19, Q-20) are 5.1 metres of

cross-laminated calcareous quartz arenites. This unit was not taken to be the Douro Formation-Somerset Island Formation contact as the influx of quartz appeared to be an isolated event and there was no tendency towards planar beds above these units (Q-19, Q-20).

Thorsteinsson and Tozer (1963) and Miall (1969) recorded numerous sandstone units in the Read Bay and Peel Sound Formation transitional sequence on eastern Prince of Wales Island. In Section PQ, approximately 60 metres of calcareous and dolomitic quartzose siltstones comprise the upper part of the Somerset Island Formation. This was the only section in which a significant thickness of clastic rocks occurs.

In other measured sections, petrographic work showed most of the arenaceous units to be quartzose dolarenites and dolosiltites rather than quartz arenites. With the exception of Section PQ, dolomitic and calcareous quartz arenites are minor constituents of the Somerset Island Formation. Miall *et al.* (1978, p. 184) did not assign this transitional sequence on eastern Prince of Wales Island to the Somerset Island Formation because of the differences in lithologies between Somerset and Prince of Wales islands. Field work on eastern Prince of Wales Island contradicts this statement, hence the introduction of Somerset Island Formation nomenclature on this island. Miall (personal communication, 1983) agreed that this revision of the nomenclature on eastern Prince of Wales Island is valid.

The top of the Somerset Island Formation is placed at the base of the lowest stratigraphic occurrence of a massive, red weathering conglomerate, conglomeratic sandstone or red weathering lithic sandstone. The upper part of the Somerset Island Formation is commonly recessive and the Somerset Island Formation-Peel Sound Formation contact is covered at many localities. A massive, red weathering conglomerate or conglomeratic sandstone occurs stratigraphically above these covered intervals and is assigned to the Peel Sound Formation. The Somerset Island Formation on eastern Prince of Wales Island has a maximum thickness 257 metres (Table 2.1).

The fauna of the Somerset Island Formation is notably less diverse than that of the Douro Formation. Rhodoliths are abundant in the lower part of the formation but decrease in abundance up-section. Oncoliths, hemispherical stromatolites, gastropods, ostracods, stromatoporoids, brachiopods, corals, and ostracoderms occur throughout the formation.

On eastern Prince of Wales Island, the Somerset Island Formation is late Ludlovian in age (Fig. 2.4). This age assignment is based on conodont faunas collected from the lower part of the formation during this study (Appendix II) and a late Ludlovian age assigned to the lower part of the lower member of the Peel Sound Formation by Thorsteinsson (1980). The upper part of the Somerset Island Formation on eastern Prince of Wales Island is usually recessive or

dolomitic and no conodont faunas were obtained.

Miall *et al.* (1978) assigned a late Ludlovian age to the base of the Somerset Island Formation on Somerset Island and a Pridolian age to the the rest of the sequence whereas Thorsteinsson (1980) assigned a late Ludlovian age to the whole of the Somerset Island Formation. The lower member of the Peel Sound Formation is late Ludlovian to early Pridolian (Miall, 1970a; Thorsteinsson, 1980). Miall *et al.* (1978) assigned a early Pridolian age to the lower member of the Peel Sound Formation on eastern Prince of Wales Island and equated the strata chronologically with the Somerset Island Formation on Somerset Island.

Peel Sound Formation

The Peel Sound Formation was named by Thorsteinsson and Tozer (1963) for a red bed sequence consisting of siltstones, sandstones, and conglomerates with calcareous and dolomitic beds near the base. The type section is on the eastern flank of the Cape Anne Syncline on northwestern Somerset Island. Two units were recognized; a lower sandstone unit and an upper conglomerate unit (Thorsteinsson and Tozer, 1963).

Miall (1969, 1970a) applied the terms, lower and upper member respectively, to what had previously been referred to as the sandstone and the conglomerate units. The lower member consists of interbedded limestone, dolostone, sandstone and oligomict conglomerate (Miall, 1969, 1970a,

1970b). Sandstone is the dominant lithology and is generally calcareous or dolomitic (Thorsteinsson, 1980). The upper member is delineated by: 1) a change in clasts from carbonates to igneous and metamorphics, 2) an increase in maximum clast size and 3) conglomerate becoming the only rock type (Miall, 1969, 1970a, 1970b). Thorsteinsson (1980) separated the lower and upper members of the Peel Sound Formation by a regional unconformity.

The contact with the underlying Douro Formation is conformable and transitional with the base of the formation arbitrarily placed at the first appearance of a red siltstone (Thorsteinsson and Tozer, 1963). Miall (1969) placed this contact at the first important clastic horizon, either the red siltstone of Thorsteinsson and Tozer (1963) or a thin pebble conglomerate. Miall and Kerr (1977) redefined the base of the Peel Sound Formation on Somerset Island as the first appearance of a cross-bedded, red sandstone. On Prince of Wales Island, Thorsteinsson (1980) placed the base of the Peel Sound Formation at the lowest stratigraphic occurrence of a planar-bedded quartzose sandstone. In this study, the base of the Peel Sound Formation is sharp, conformable and placed at the lowest stratigraphic occurrence of a thick, red weathering conglomerate or conglomeratic sandstone.

Two members of the Peel Sound Formation are still recognized on eastern Prince of Wales Island. The lower member equates with the upper part of the lower member of

the Peel Sound Formation of Thorsteinsson (1980). Strata previously assigned to the lower part of the lower member of Thorsteinsson (1980) have been reassigned to the Somerset Island Formation. The upper member of the Peel Sound Formation as described by Thorsteinsson (1980) remains unchanged.

The lower member of the Peel Sound Formation is late Ludlovian to early Pridolian in age (Miall, 1970a; 1970b; Thorsteinsson, 1980). Thorsteinsson (1980) assigned the upper member of the Peel Sound to the late Lochkovian on the basis of the absence of late Pridolian to middle Lochkovian fauna. Miall *et al*. (1978) assigned the upper member to the late Pridolian and early Gedinnian (Fig. 2.4).

III. CLASSIFICATION OF ROCK TYPES

A. Introduction

The strata of this study are lithologically diverse and some discussion of the terminology and classification of these rock types is required. A twofold classification scheme was applied to the rocks of this study. The first is a compositional classification based on the relative percentages of three end members (quartz, calcite and dolomite) irrespective of their distribution in the rock. The second is a textural classification based on the distribution of the end members in the rock once the composition has been determined. The compositional and textural classifications may overlap as a texture may encompass more than one composition and visa versa. These lithologies (rock types) are purely descriptive terms, based on petrographic and field analysis, and no genesis is implied.

Lithology, when used in conjunction with sedimentological and paleontological criteria, constitutes a lithofacies. De Raaf *et al*. (1965) defined lithofacies as the lithological, structural and organic aspects of a rock unit observable in the field and which ultimately give an environmental interpretation to the rock unit. This definition is expanded to include petrographic data.

B. Compositional classification

Initially, all rock units were classified solely on the basis of the relative percentages of three dominant end members: calcite, dolomite and detrital quartz grains. Secondary chalcedony was delineated from detrital quartz and classified as an accessory mineral. Other accessory minerals include collophane, feldspar, hematite, pyrite, muscovite, azurite and malachite. An accessory mineral, as defined herein, refers to any constituent that comprises less than 10% of the rock and in some situations may include the end members. Typically, accessory minerals constitute less than 2% of a rock.

Recognition of compositional rock types was for the purposes of establishing generalized lithologies that may be applied to a more descriptive textural classification. Fifteen compositional rock types are recognized (Fig. 3.1).

Pure rock types

Pure rock types are composed of greater than 90% of any one end member and less than 10% of any remaining constituent, whether that constituent is either or both of the other end members, an accessory mineral or any combination of the above. Three pure rock types are recognized:

1. Limestone: comprised of greater than 90% calcite,
2. Dolostone: comprised of greater than 90% dolomite (note dolomite refers to the

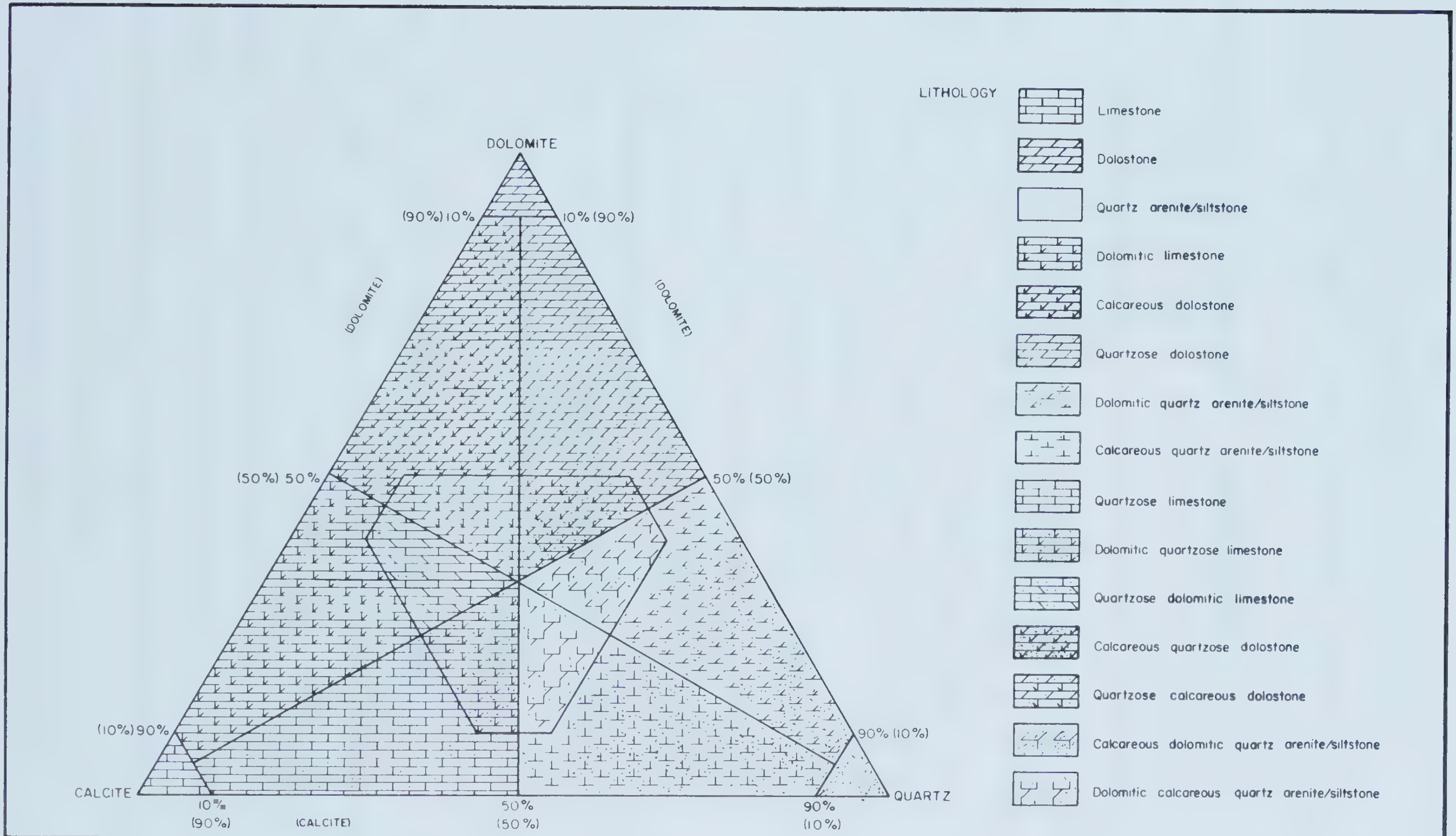


FIG. 3.1. Compositional classification of lithology types.

mineral only) and

3. Quartz arenite/siltstone: comprised of greater than 90% detrital quartz grains.

Impure rock types composed of two end members

In these rock types, the dominant end member constitutes greater than 50% but less than 90% of the rock. The modifying end member constitutes greater than 10% but less than 50% of the rock and the remaining end member and/or accessory minerals constitute less than 10% of the rock. In this group, six rock types are recognized:

1. Dolomitic limestone,
2. Quartzose limestone: quartzose denotes either or both quartz silt- or sand-sized grains,
3. Calcareous dolostone,
4. Quartzose dolostone,
5. Dolomitic quartz arenite/siltstone and
6. Calcareous quartz arenite/siltstone.

Impure rock types composed of three end members

In these rock types, all three end members are present and constitute greater than 10% of the rock with the dominant end member constituting greater than 33% but not exceeding 50% of the rock. The order of naming indicates the ascending percentages of the respective end member. In this group, six rock types are recognized:

1. Dolomitic quartzose limestone,

2. Quartzose dolomitic limestone,
3. Calcareous quartzose dolostone,
4. Quartzose calcareous dolostone,
5. Calcareous dolomitic quartz arenite/siltstone and
6. Dolomitic calcareous quartz arenite/siltstone.

C. Textural classification

To further describe the lithologies, they are classified according to the distribution and nature of the end members within the rock type. For example, calcite may occur as bioclasts, intraclasts, micrite, sparite, or any variation of the forementioned in a single sample. Compositionally, the rock may be a pure end member but texturally a spectrum of lithologies could be represented. Each of the end members will be discussed along with the textural and compositional variations found in the Upper Silurian strata on eastern Prince of Wales Island.

Limestones

The textural classification for limestones is modified from Folk (1959, 1962, 1974). This classification (Fig. 3.2) incorporates aspects of both the compositional and textural spectrums of Folk (1959, 1962, 1974) and modifies these to describe the rock types present in this study. Only two allochems, bioclasts and peloids, occur in sufficient abundance to warrant use as modifiers. Ooids occur as isolated grains in a single sample and intraclasts do not

FOLK (1974)

	LIME MUD MATRIX					SPAR CEMENT		
% Allochems	0-1 %	1-10 %	10-50 %	OVER 50%	SPAR & LIME MUD	SORTING POOR	SORTING GOOD	ROUNDED & ABRADED
Rock Terms	MICRITE & DISMICRITE	FOSSILI-FEROUS MICRITE	SPARSE BIOMICRITE	PACKED BIOMICRITE	POORLY-WASHED BIOSPARITE	UNSORTED BIOSPARITE	SORTED BIOSPARITE	ROUNDED BIOSPARITE

THIS STUDY

	MICRITE MATRIX				SPARITE CEMENT	
Percent Bioclasts	0 %	1-10 %	10-50%	OVER 50%	GRAIN SUPPORTED	ORGANIC FRAMEWORK
Lithology	MICRITE	FOSSILI-FEROUS MICRITE	SPARSE BIOMICRITE	PACKED BIOMICRITE	SKELETAL CALCARENITE	BIOLITHITE
Percent Peloids	0-10%		OVER 10%			
Lithology	MICRITE		PELMICRITE			

FIG. 3.2. Textural classification of limestones.

comprise greater than 10% of any rock.

Micrites: Pure micrites are relatively rare in the strata of this study. The micrites are generally dark grey, thinly laminated to massive (Table 3.1) and form units usually less than metre thick. Typically, the lower contact with the underlying unit is sharp and the upper contact is gradational. A field criteria used to identify micrites was the presence of a conchoidal fracture on a fresh surface.

Fossiliferous micrites and sparse biomicrites: These two lithologies, with no subsequent modifier such as quartzose or dolomitic, are rare. The rocks are generally medium grey to brown-grey, massive and form units less than a metre in thickness. Bioclasts form 1-50% of the rock (Fig. 3.2). Contacts of the fossiliferous micrites and the sparse biomicrites with adjacent units are gradational. Sparite may occur as a fracture fill, cement, neomorphic texture or as a geopetal fabric in association with bioclasts.

Packed biomicrites: Packed biomicrites occur as lenses in other lithologies. Bioclasts form more than 50% of the rock (Fig. 3.2) and the bioclast population can be monospecific or of a diverse assemblage.

Pelmicrite and (bio)pelmicrite: Micrites which contain greater than 10% peloids are given the modifier *pel* (Fig. 3.2). The term *peloid* refers to an allochem formed of microcrystalline or cryptocrystalline material irrespective of size or origin (McKee and Gutschick, 1969). Folk (1959, 1962, 1974) defined a similar term *pellet* for homogenous

DESCRIPTION	CHARACTER
No bedding apparent	Massive
Very thickly bedded	Thicker than 1 m
Thickly bedded	30-100 cm
Medium bedded	10-30 cm
Thinly bedded	3-10 cm
Very thinly bedded	1-3 cm
Thickly laminated	0.3-1 cm
Thinly laminated	Thinner than 0.3

TABLE 3.1. Scale of stratification thickness (from Ingram, 1954, *in* Blatt *et al.*, 1980, p. 128).

aggregates of microcrystalline calcite, ranging from 30-200 microns in diameter that can be biogenic in origin or a form of recrystallization. The peloids are less than 2 mm in diameter. The term *peloid* is preferred over *pellet* as the latter term has assumed an implied biogenic origin rather than being a nongenetic term. They can be recognized in the field but are most evident in thin section.

Where bioclasts are also present but comprise a smaller percentage of the rock than the peloids, the rock is a biopelmicrite and visa versa. Typically, the pelmicrites are medium brown-grey on a fresh surface. The upper and lower contact of the pelmicrite units are gradational and no mean thickness for these units was established. Pelmicrites without a modifier such as dolomitic are rare.

Skeletal calcarenites: Rocks in which the allochems are recognizable as being bioclastic in origin, are grain supported and have a sparry calcite cement are termed *skeletal calcarenites*. These rocks occur as lenses in other rock types or as distinct field units. Commonly, the bioclast population in the lenses is dominated by crinoids whereas the field units are composed of a diverse assemblage of allochems. The units may be massive with no bedding character or rubbly weathering, several metres thick, show sharp lower contacts and gradational upper contacts with other units.

Dolomitic limestones: Dolomitic biomicrites and pelmicrites are common lithologies. Dolomitic calcarenites are less common. The dolomite can occur as interbeds or as irregular stringers in the limestone. Typically, the dolomitic limestones are medium brown-grey on fresh surfaces and weather light to medium grey. The dolomitic limestones are massive to rubbly; the bedding character and thickness of the units being highly variable. The contacts with adjacent units are generally gradational and based primarly on weathering profile.

Quartzose limestone, dolomitic quartzose limestone and quartzose dolomitic limestone: These three rock types represent the introduction of quartz detritus into any of the forementioned limestone types.

The quartz grains are variable in distribution, usually dispersed throughout the rock and are rarely clustered. The

sand-sized grains are generally sub-rounded, moderately sorted and show uniform extinction under cross-nichols. In some rocks, the grains are highly fractured and show undulose extinction. The silt-sized quartz grains are generally sub-angular and show a lesser degree of sorting than the sand-sized grains.

Dolostones

Dolostones are classified according to their grain size, regardless of genesis (Table 3.2). Folk (1974, p. 167) recognized two distinct types of carbonate particles; transported constituents to which he applied the Wentworth scale and authigenic constituents to which a crystalline size (Table 3.2). A range of petrographic textures occurs in the dolostones but in the field only the grain size is apparent and the dolostones were classified as dololutites, dolosiltites, dolarenites and dolorudites (Table 3.2).

Dolosiltites/dolarenites and quartzose dolosiltites/dolarenites: These two dolostone types represent the dominant dolostone types in the Upper Silurian stata on eastern Prince of Wales Island. The dolostones are commonly light green-grey to light blue-grey on fresh exposure and weather the same. The units are usually thinly laminated to thinly bedded (Table 3.1). Generally, both the lower and upper unit contacts are gradational. The delineation into field units is based primarily on bedding thickness. Rarely do these units exceed 0.5 metres in

GRAIN SIZE (NO GENESIS IMPLIED)
THIS STUDY

Lower limit	(Transported constituents Folk (1974))	(Authigenic constituents Folk (1974))
2.00 mm	Dolorudite/calcirudite	Very coarsely crystalline
1.00 mm	Very coarse dolarenite/calcarenite	
0.50 mm	Coarse dolarenite/calcarenite	Coarsely crystalline
0.25 mm	Medium dolarenite/calcarenite	
0.062 mm	Fine dolarenite/calcarenite	Medium crystalline
0.031 mm	Coarse dolosiltite/calcisiltite	Finely crystalline
0.016 mm	Medium dolosiltite/calcisiltite	
0.004 mm	Fine dolosiltite/calcisiltite	Very finely crystalline
0.001 mm	Dololutite/calcilutite (micrite)	Aphanocrystalline

TABLE 3.2. Grain size classification of dolostones and limestones.

thickness.

Petrographically the dolosiltites are xenotopic to idiotopic mosaics of dolomite grains, the dominant texture is xenotopic. The quartz grains are petrographically similar to quartz grains in the quartzose limestones. Silt-sized grains are more prevalent than sand-sized grains. Generally, the quartz grains are slightly larger than the dolomite grains. This size difference may be related to the genesis of the rock type.

Calcareous dolostones: Texturally this rock type is very similar to the dolosiltites/dolarenites but contains calcite as either calcareous grains, rare bioclasts, or a cement. Dolosilts are the dominant particle size. The rock is generally light green to light grey-green, has no bedding structure, a sharp basal contact and a gradational upper contact.

Quartzose calcareous dolostones: These rock types are generally dolostones with thin limestone interbeds or dolostones with isolated calcareous lumps which are commonly bioclasts. Calcite may also occur as a cement. The quartz can occur in either the dolostone or limestone portion and tends to decrease up-unit. Dolosilts are the dominant particle size. The rock types are generally medium grey to dark green-grey and commonly occur as transitional units with dolomitic limestones. The units are massive to thin bedded and the lower and upper unit contacts are usually gradational. This rock type generally weathers recessive

which may be a factor in it being volumetrically unimportant in the strata of this study.

D. (Dolomitic and calcareous) quartz arenites/siltstones

All the textural types in the quartz arenites/siltstones can be discussed together. The only variables are the type of cementation and the presence or absence of bioclasts and/or lithoclasts. The sand-sized grains are generally well rounded and well sorted whereas the silt-sized grains are sub-angular to sub-rounded and moderately sorted. The grains usually show uniform extinction and commonly show some corrosion along the grain margins. In a carbonate sequence, the units show both sharp lower and upper contacts whereas in a clastic sequence the contacts are generally gradational. The units generally lack any bedding or internal structure. In only Section PQ were these rock types volumetrically important. These rock types are significant in that they commonly mark the boundary between the Douro and Somerset Island formations.

IV. ALGAL STRUCTURES

A. Introduction

Algal structures are abundant, diverse and occur throughout the Upper Silurian strata on eastern Prince of Wales Island. Algae representative of the phyla Schizophyta and Rhodophycophyta occur in all three formations of this study. The Cape Storm Formation is dominated by Schizophyta and the Douro Formation by Rhodophycophyta, whereas the Somerset Island Formation contains both phyla in abundance. The algal phylum Chlorophycophyta occurs only in the upper Douro Formation.

The variations in morphology and diversity of algal types, in association with the host lithology, posed the problem of morphogenesis and the interaction of abiotic and biotic factors on morphology. The study of stromatolites is essentially one of morphogenesis but this argument may be extended to include porostromate and coralline algal structures, both of which show a diversity of morphologies.

Two "Families" of Schizophyta (Pia, 1927, cited in Johnson, 1961) are recognized, Porostromata which is characterized by a filamentous microstructure and Spongiostromata which is characterized by a cryptalgal fabric. The two "Families" do not occur together in any stratum of this study. Although a variety of spongiostromate structures occur in this strata, hemispherical stromatolites of the Cape Storm and Somerset Island formations dominate

and will be discussed in greater detail.

The most common porostromate algal structures are the oncoliths in the Somerset Island Formation. These nodular structures are not to be confused with coralline algal nodules (rhodoliths) which can occur in the same stratum. Both oncoliths and rhodoliths have spheroidal and irregular shapes which may be useful for paleoenvironmental interpretation. Porostromate algae also occur in hemispherical structures composed of *Girvanella*, *Sphaerocodium* and the problematic algae *Wetheredella*.

In the Upper Silurian strata on eastern Prince of Wales Island, the phylum Rhodophycophyta is represented by the genus *Solenopora*, of which two species are recognized. The rhodoliths, composed of *Solenopora filiformis*, have the longest stratigraphic distribution of the algae of this study. A second species, *Solenopora* sp., occurs as small bioherms in the Somerset Island Formation. *Solenopora* occurs in a variety of lithologies and with different faunal associations suggesting a spectrum of paleoenvironments.

This study of algae was initiated in response to the abundance and diversity of algal structures in the Upper Silurian strata on eastern Prince of Wales Island. The purposes of studying the algae were twofold: 1) record the diversity in form of specific algal groups and 2) relate this diversity and/or the occurrences of algal structures to the stratigraphic setting in an attempt to use algae as a paleoenvironmental tool.

The taxonomy of algae is in constant flux and controversy, hence the taxonomy of the algae found during this study will be reviewed. Diagnosis and discussion are based solely on the literature, whereas the descriptions and remarks are based on the specimens found during this study. This format is not intended to resolve the problems in taxonomy but to clarify discussions in this chapter.

B. Schizophyta

Pia (1927, cited in Johnson, 1961) defined two "Families" (=Sections) of Schizophyceae with uncertain affinities; Spongiostromata for fossil forms presumably built by algae and showing no filamentous or cellular microstructure, but develop colonies having a constant shape and Porostromata for all fossil algae characterized by a microstructure of calcified filaments. Monty (1981), however, recognized that neither microstructure is exclusive to Porostromata nor Spongiostromata. In this study, the term *microstructure* is restricted to porostromate algae with a filamentous microstructure. The term *cryptalgal fabric* is restricted to spongiostromate microstructure.

The "Family" *Spongiostromata* was divided into two groups; Stromatolithi (=stromatolite of Kalkowski, 1908) for the various shaped, laminated algal forms growing attached to the substrate and Oncolithi (=oncolite of Kalkowski, 1908) for calcareous nodules growing free from the substrate. On the basis of algal structures found in the

Upper Silurian strata on eastern Prince of Wales Island and the cryptalgal microstructures of Riding (1977a) and Monty (1981), the "Family" *Spongiostromata* will be expanded to include thrombolites, algal laminites and fenestrae. All the oncoliths (=oncolites, =oncoids) observed during this study have a filamentous microstructure and are, thus, assigned to Porostromata.

Three genera of filamentous blue-green algae are present in the strata of the study area and are assigned to the "Family" Porostromata. *Girvanella* and *Sphaerocodium* are of an algal affinity, whereas *Wetheredella* is tentatively assigned to the schizophytes. All the porostromate algae, of this study, are found in structures more adequately described in the spongiostromate nomenclature of stromatolites and oncoliths. The presence of a filamentous microstructure takes precedence over external morphology and the porostromate stromatolites and oncoliths are retained in Porostromata.

The confusion and debate over taxonomic nomenclature is endless and appears to be modified for the purposes of individual studies. Riding (1977a), in an attempt to clarify the issue, introduced the term *skeletal stromatolite/oncolith* for algal structures in which the organisms responsible for their formation are usually preserved as calcified fossils and the term *nonskeletal stromatolite/oncolith* in which the micro-organisms are not calcified. The adjective *skeletal* (Riding, 1977a, p. 58) may

be used in reference to either generic or geometric descriptions of stromatolites or oncoliths. This adjective appears to transcend "family" boundaries; a geometric classification is applied to the algal structures of this study that lack in any other criteria such as microstructure for classification. Monty (1981) recognized porostromate stromatolites and oncoliths and suggested they mark an evolutionary step in the uppermost Precambrian when calcified filaments first appear.

In this study, Spongiostromata and Porostromata are retained but modified to more accurately describe the specimens found throughout this study. In order to clarify further discussions, more terms need to be defined. Thus, stromatolites refers to spongiostromate algal structures characterized by the presence of laminae and attachment to the substrate. Classification of these stromatolites is based on the geometry of their external morphology. Oncolith refers to laminated or nonlaminated porostromate algal structures comprised predominantly of *Girvanella* filaments. This growth form is nodular and usually forms detached from the substrate.

Porostromata

Phylum SCHIZOPHYTA (Falkenberg) Engler, 1892

(Blue-green algae)

"Section" (Family) POROSTROMATA Pia, 1927

Genus SPHAEROCODIUM Rothpletz, 1890

Type species

Sphaerocodium bornemanni Rothpletz (1890, p. 9).

Synonyms: *Rothpletzella* Wood, 1948

Coactilum Maslov, 1956

Diagnosis

Thallus exhibits a variety of growth habits, from encrusting forms a few layers of filaments thick, to columnar masses intergrown with other organisms. Individual tubular filaments are nonseptate and elliptical in transverse section, being flattened against the surface of attachment. Tangential sections of the filaments show a distinct branching habit with the branches diverging in a fanlike pattern while maintaining contact along the inner walls. The filaments have distinct walls composed of microcrystalline calcite.

Discussion

The algal genus *Sphaerocodium* was proposed by Rothpletz (1890) for *S. bornemanni* from the Triassic of the eastern Alps. In 1908 and 1913, Rothletz named *S. gotlandicum* and *S. munthei* from the Silurian of the Baltic region. Pia (1927) and Wood (1948), however, believed *Sphaerocodium* to be a complex of several different algae. The assignment of *Sphaerocodium* to the algae was questioned by Riding (1977b, p. 94) but no other affinity was suggested.

Wood (1948) re-examined the type material of *Sphaerocodium* and recognized the symbiotic intergrowths of two algal genera and an encrusting foraminifera (*Wetheredella silurica* Wood, 1948). Consortiums of foraminifera and schizophytes were also documented by Johnson (1950) and Peryt and Peryt (1975). Wood (1948) proposed the algal genus *Rothpletzella* for the larger branching filaments and placed the smaller non-branching filaments in the genus *Girvanella*. Johnson and Konishi (1959), Johnson (1961), Johnson (1964) and Copper (1976) recognized *Rothpletzella* as a valid genus. Wray (1967a, 1967b, 1977), however, suggested that *Sphaerocodium* had priority under the Code of Botanical Nomenclature and that *Rothpletzella* and *Coactilum* are thus synonyms of *Sphaerocodium*.

Wood (1948) proposed the genus *Wetheredella* for an encrusting foraminifera which commonly occurs as intergrowths with *Spongiostromata*, *Girvanella* and

Rothpletzella. The semicircular tubes are unchambered and generally flattened parallel to the surface of attachment. Individual tubes range from 30-150 microns in diameter and have a distinct wall structure composed of radial fibrous calcite.

Copper (1976) supported Wood's (1948) taxa but considered *Wetheredella* to be a skeletal blue-green algae rather than a foraminifera. Loeblich and Tappan (1964) did not recognize *Wetheredella* as a foraminifera but as possibly an algae, whereas Johnson (1964) and Wray (1967a) recognized *Wetheredella* as a foraminifera. *Wetheredella silurica* Wood, 1948 and *Sphaerocodium munthei* (=*Rothpletzella munthei*) Rothpletz, 1913 are similar and may be considered synonymous (Copper, 1976; Edhorn, 1979), supporting the argument for assigning *Wetheredella* to the schizophytes. *Wetheredella* was recognized in rocks from the Douro Formation on Somerset Island (Narbonne, 1981; Narbonne and Dixon, 1984) and was tentatively considered to be an algae (Narbonne, 1981, p. 98).

In this study *Sphaerocodium* is considered the valid taxonomic name. *Wetheredella* is also retained and tentatively assigned to the schizophytes.

Sphaerocodium sp.

(Plate 1, Figs. 1, 2 and 3)

Description

Thallus shows an encrusting habit with multiple layers of filaments forming masses up to 1.5 mm thick. In transverse section, the filaments (Plate 1, Figs. 1, 2 and 3) show the characteristic beaded appearance of *Sphaerocodium*. Individual filaments are 30-40 microns high and 40-60 microns wide in transverse section, the filaments being flattened against the surface of attachment.

In tangential section, the filaments are nonseptate and branch dichotomously at least twice, giving the characteristic fanlike pattern of the filaments (Plate 1, Fig. 1). Wall thickness is 4-8 microns and is best defined in tangential section; the filament walls are poorly preserved and rarely evident in transverse section. A maximum filament diameter of 110 microns was measured in tangential section. No branching interval was observed and equatorial flanges are absent. Anhedral calcite fills the cells.

Wetheredella? sp.

(Plate 1, Figs. 2 and 3; Plate 2, Fig. 4)

Description

Thallus are encrusting, generally consisting of single layers of filaments. In transverse section, the tubular filaments are 300-350 microns high and 400-450 microns wide, being flattened against the surface of attachment. (Plate 1, Figs. 2 and 3). Filament walls are 40 microns thick, well defined and show a radial fibrous calcite structure as described by Wood (1948). In longitudinal section, the filaments tend towards a radial pattern and are nonseptate. Branching is absent and no perforate structure of the walls was observed. The serial chain of filaments appear to share a common wall (Plate 1, Figs. 2 and 3). This common wall is not always present giving a bilobate or multilobate appearance to the filaments. *Wetheredella* shows a relatively low degree of organization in the columnar complexes (Plate 3, Figs. 1 and 2), the filaments becoming random in size, shape and orientation in some layers.

Remarks

Sphaerocodium sp. only occurs in one sample from unit M-56, occurring as intergrowths with *Girvanella* sp. and *Wetheredella* sp. (Fig. 4.1) to form columnar masses up to 6 cm high (Plate 4, Figs. 1 and 2). The algal columns are cylindrical in transverse view, showing a denser assemblage of filaments in the central core than on the peripheries.

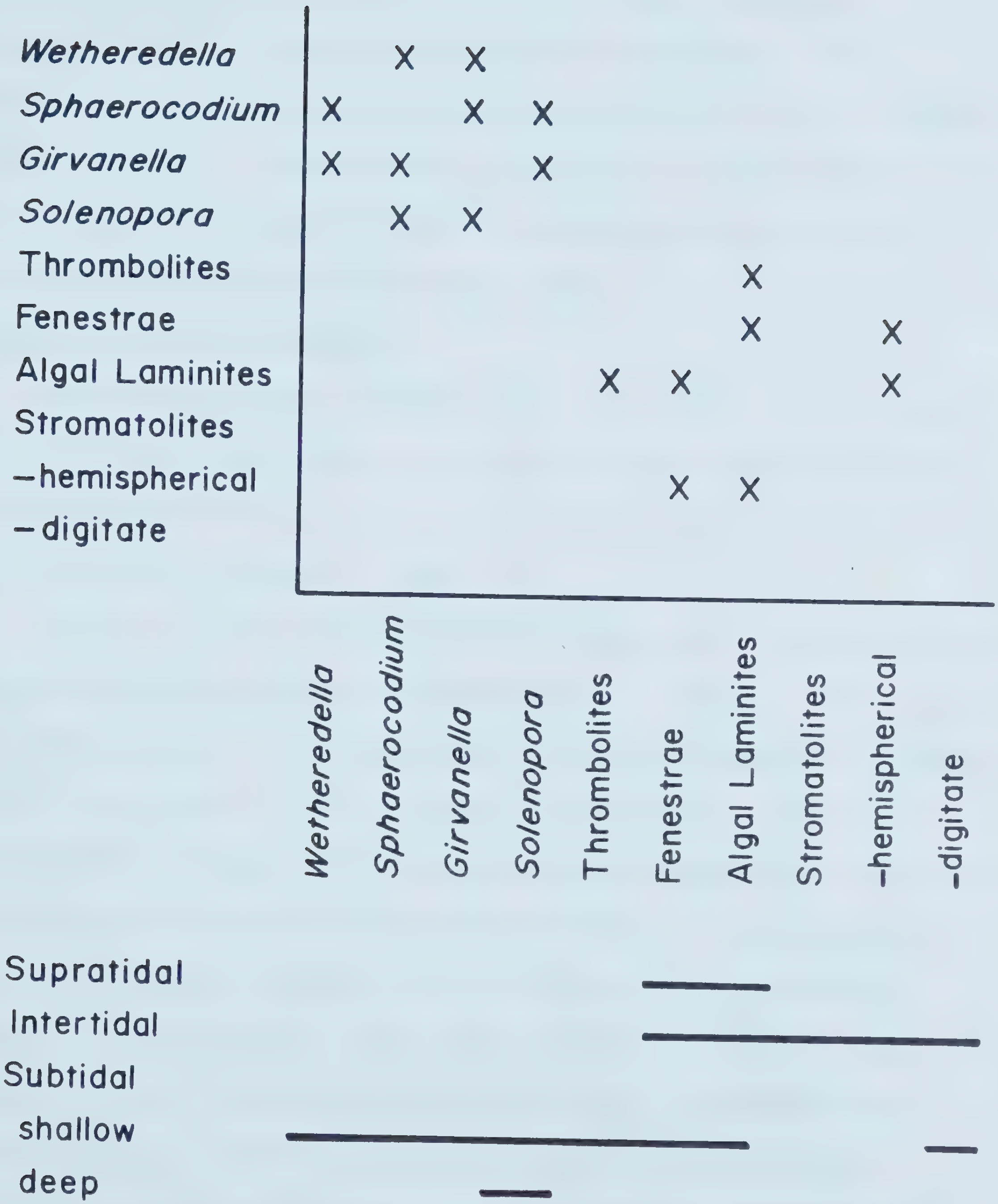


FIG. 4.1. Algal associations and bathymetric range of algae in the Upper Silurian strata on eastern Prince of Wales Island.

The columns are concentrically laminated, the laminae consisting of multiserial chains of algal filaments. *Sphaerocodium* sp. also occurs as rare filaments with *Girvanella*, as encrusting masses on coralline algal nodules (Plate 3, Figs. 1 and 2). The specimen of *Sphaerocodium* sp. is similar to *S. gotlandicum* Rothpletz but because this is one of the species of taxonomic debate in the literature, no species will be assigned.

Wetheredella sp. occurs only in two samples from units M-56 and M-73. The first occurrence is described above and the second occurrence is as intergrowths with *Girvanella* sp. in an oncolith (Plate 2, Fig. 4).

There is some doubt as to the taxonomic affinity of the specimens identified as *Wetheredella* sp. during this study. The size range of the individual filaments observed by Wood (1948) and Copper (1976) is typically 30-150 microns, whereas the diameter of the individual filaments in M-56 is 300-450 microns. The filaments identified as *Wetheredella* sp. are similar in appearance to those figured by Wood (1948), Johnson (1964) and Wray (1967a) but unlike that of Copper (1976). The taxonomic description of *Wetheredella tumulus* Copper (1976) made no reference to the radial fibrous calcite wall structure as described by Wood (1948). This wall structure is well preserved in the specimens of *Wetheredella* sp. from units M-56 and M-73. The *Wetheredella* of Johnson (1964, Plate 28, Fig. 4) and Wray (1967a, p. 65, Fig. 7) are similar in size and appearance to those

identified in this study.

Narbonne (1981) and Narbonne and Dixon (1984) noted similar organisms in association with *Girvanella*, *Sphaerocodium* and *Wetheredella* but tentatively assigned them to the coral genus *Aulopora*. *Aulopora* may occur as encrusting biserial or multiserial rows of laterally continuous corallites lacking in either septa or tabulae (Hill and Stumm, 1956, p. F472).

The large tubular filamented organism in the algal complexes of unit M-56 were identified as *Sphaerocodium* (Wray, written communication, 1984). This identification is questioned by this author and on the basis of wall structure, size of the filaments and the lack of branching of the filaments is assigned to the algal genus *Wetheredella*. Only two species of *Wetheredella* (*W. silurica* Wood, 1948 and *W. tumulus* Copper, 1976) have previously been recognized. Riding (1977b), however, refuted Copper's (1976) taxonomy and considered *W. tumulus* Copper as a synonym of *W. silurica* Wood. He also argued against assigning *Wetheredella* to the cyanophytes but left the question of affinity open. Neither description adequately describes the algae identified as *Wetheredella* sp. from the Upper Silurian strata on eastern Prince of Wales Island, suggesting that these specimens may belong to a new species.

Paleoenvironment and stratigraphic occurrence

Typically *Sphaerocodium* is found in Paleozoic reefs and bioherms as a frame-builder or as a frame-binder (Hadding, 1950; Johnson, 1964; Wray, 1967a, 1967b, 1977; Flügel and Wolf, 1969; Wray and Playford, 1970; Pitcher, 1971; Machielse, 1972; Tsien and Dricot, 1977; Narbonne, 1981; Narbonne and Dixon, 1982). Tsien and Dricot (1977) observed columnar masses of *Sphaerocodium* in the lagoonal facies in the Devonian of the Dinant and Namur basins in Belgium. Copper (1976) considered *Wetheredella* a frame-builder and frame-binder in reef and off-reef sediments in the Ordovician on Anticosti Island. The only previously recorded occurrence of *Sphaerocodium* and *Wetheredella*, from the Upper Silurian of the Arctic Lowlands, was as encrustations on lithistid sponge reefs in the Douro Formation on Somerset Island (Narbonne, 1981; Narbonne and Dixon, 1982).

On an exposed bedding plane, the columnar algal complex in unit M-56 resembles a hemispherical stromatolite. The undulating bedding plane surface has a maximum relief of 9 cm. Single hemispheres are comprised of several columnar algal complexes capped by a mat of *Girvanella* filaments (Plate 3, Figs. 1 and 2). Similar growth forms of algal complexes were documented by Wray (1967a) in the Devonian of the Canning Basin and by Copper (1976) in the Ordovician on Anticosti Island.

Associated with the algal complex in unit M-56 is a relatively diverse fauna of high spired gastropods,

orthoconic nautiloids, *Coenites*-like corals, *Atrypoidea erebus* and *Protathyris praecursor*. A dolomitic sparse biomicrite fills the spaces between the algal complexes and the interareas of the hemispherical domes. No sedimentological structures occur in this unit.

There are several lines of evidence to suggest that the algal complexes were subjected to at least intermittent wave or current action. The algal columns show disruption and fragmentation of the layers, particularly in the basal portion of the columns (Plate 3, Figs. 1 and 2). Also, the concentric laminae of the algal complexes are truncated suggesting scour along the margins of the complexes. Megafossils do not show fragmentation only disruption from life position and the high bioclastic content relative to the abundance of megafossils suggests that at least some of the bioclasts are allochthonous to the stratum. The bioclast content attests to some degree of reworking of the sediment but not to extensive winnowing as shown by the angularity of the bioclasts and the retention of a micrite matrix.

The *Solenopora* nodules (=rhodoliths) in unit M-56 show evidence of abrasion and later encrustation on the abraded filaments. Encrusting masses of *Girvanella* sp. with rare *Sphaerocodium* sp. filaments occur only on the upper side of the *Solenopora* nodules (Plate 3, Figs. 1 and 2). Encrusting *Favosites* also occur only on the upper side of the rhodoliths. This suggests a relatively stable position of the rhodoliths after initial transportation or disruption.

If the nodules were detached from the substrate and subjected to constant agitation, a more uniform concentric layer of *Girvanella* would develop, such as the Type "R" or Type "C" spheroidal structures (Fig. 4.2a) of Logan *et al.* (1964). Approximately 1.5 metres above the algal columns in unit M-56, in unit M-58, wave generated cross-laminations and ripple marks occur.

The algal complex in unit M-56 probably developed in a moderate energy, shallow subtidal environment, near or above wave base. This is in contrast to the common association of *Sphaerocodium* with Paleozoic reefs (*op. cit.*, p. 56). The lower part of the Somerset Island Formation is considered to be shallow subtidal to low intertidal in origin since it contains; interclast breccias (Plate 5, Fig. 2; Plate 6, Fig. 4), desiccation polygons (Plate 6, Figs. 1 and 3), algal stromatolites (Plate 7, Figs. 1 and 2), ripple marks (Plate 6, Fig. 3), cross-laminated calcarenites and calcisiltites and a low diversity fauna.

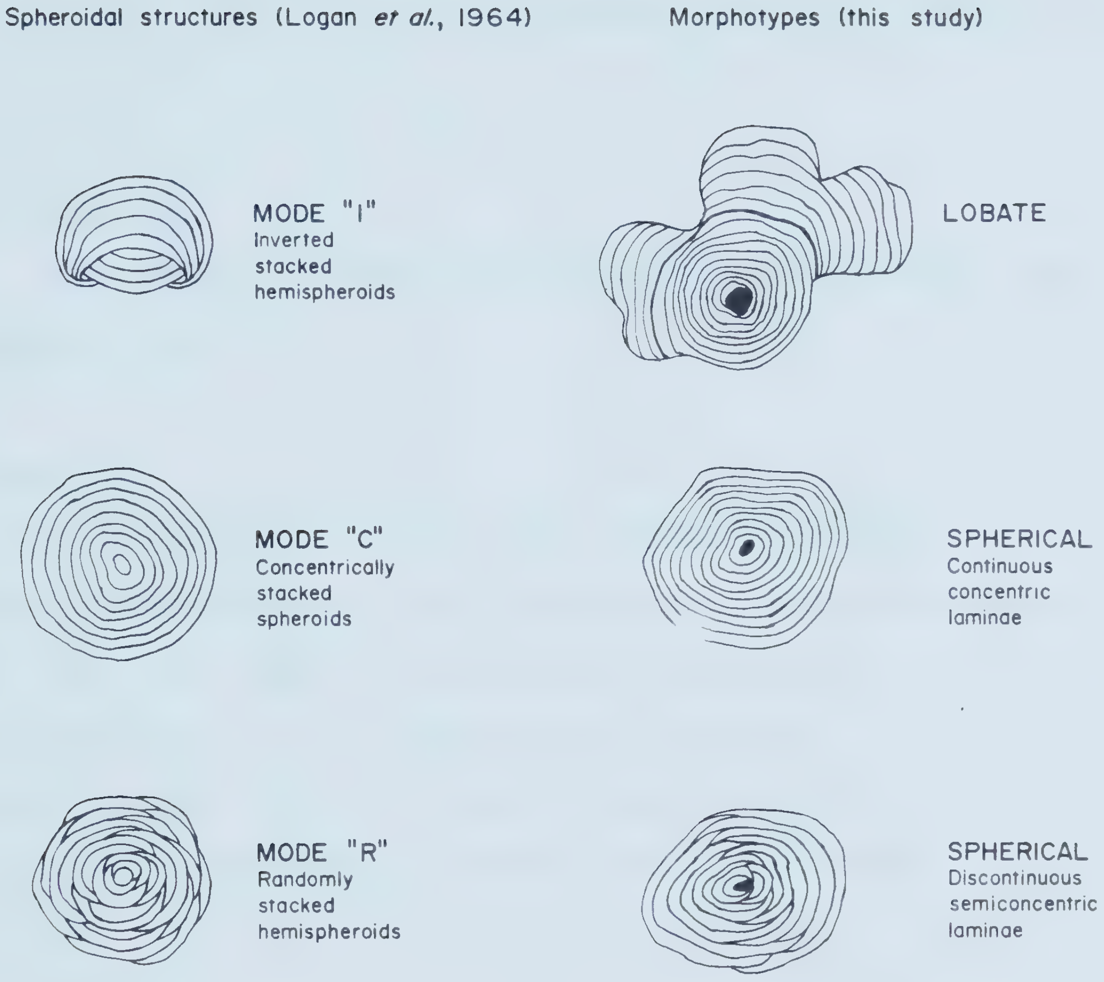


FIG. 4.2a. Comparism of spheroidal structures (Logan *et al.*, 1964) versus oncolith morphotype.

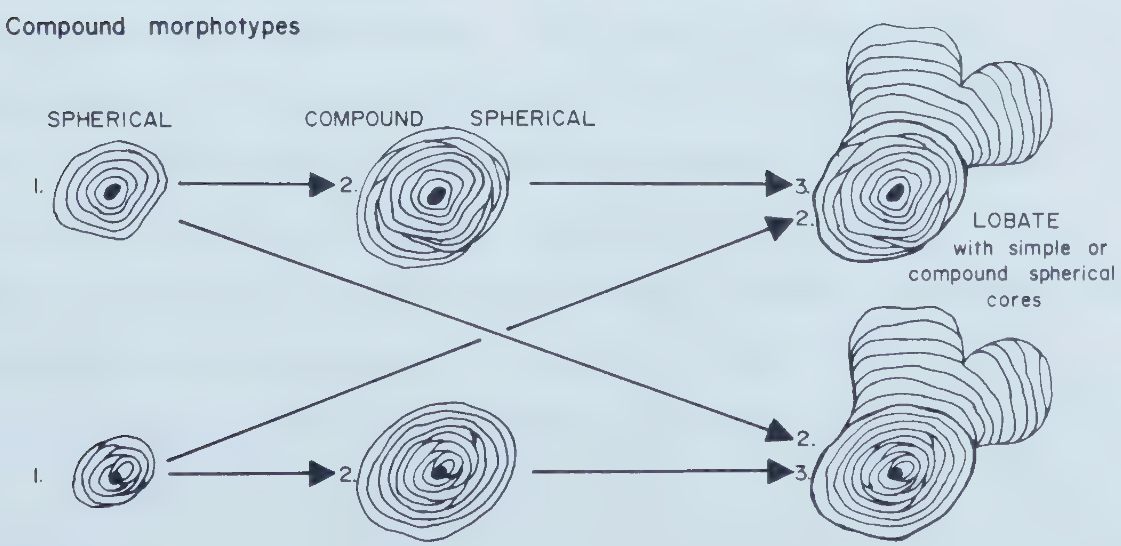


FIG. 4.2b. Growth stages of compound oncoliths.

Genus GIRVANELLA Nicholson and Etheridge, 1880

Type species

Girvanella problematica Nicholson and Etheridge (1880, Plate 9, Fig. 24).

Diagnosis

Thalli consists of flexuous tubular filaments of uniform diameter with thick, well defined calcareous walls. The tubes are simple cylinders without cross-partitions or perforations of the filament walls. Branching is rare. Thalli may be loose but usually occur as clusters twisted together to form nodular and/or encrusting masses.

Discussion

Girvanella has been separated into species primarly on the basis of the diameter of the tubes and wall thickness. When present, the character of branching may be used in speciation (Johnson and Konishi, 1959, p. 56). Outside diameters average 10-30 microns but may be as large as 100 microns (Wray, 1977, p. 36). *Girvanella* usually had an encrusting habit. Johnson and Konishi (1959) considered the taxonomy of *Girvanella* to be confusing and that careful restudy may discard many of the species.

Girvanella problematica Nicholson and Etheridge, 1880

Nicholson and Etheridge, 1880

(Plate 8, Figs. 3, 4, 5 and 6)

Description

Thallus consists of small, nonseptate, tabular filaments with a well defined wall. Outside diameters range from 13-17 microns with a mean diameter of 16 microns. Wall thickness is 2-3 microns. Thalli commonly consist of intricate bundles of interwoven filaments, forming masses less than 500 microns in size. Filaments can occur singularly. Branching was not observed.

Remarks

G. problematica is recognized by its growth habit and diameter of the filaments. The characteristics of *G. problematica* are markedly different than *Girvanella* sp. which occurs in the oncoliths. The masses (buttons) of *G. problematica* are evident only in thin section. Within the masses, the filaments are loosely felted and can be randomly orientated or subparallel to adjacent filaments. Typically the filaments occur as intergrown masses (Plate 8, Figs. 3, 5 and 6) but some filaments can extend outwards from the masses (Plate 8, Fig. 4). Edhorn (1979) noted similar growth habits in *G. problematica* from the Lower Cambrian and attributed these growth habits to mobility of the filaments as well as the ability of the filaments to expand and contract. In the masses of *G. problematica*, rows of circular

cell structures occur (Plate 8, Figs. 4 and 6). Edhorn (1979) recognized similar rows of circular cell structures in Lower Cambrian *G. problematica* and considered them to be a reproductive stage of the algae.

The filaments do not appear to be effected by recrystallization; well preserved *G. problematica* masses occur in sparry calcite zones in the biomicrite. In these zones, the filaments were conspicuous by their well-defined wall. The masses do not appear to be encrusting as is the case with *Girvanella* sp. in the oncoliths and the algal complexes.

Paleoenvironment and stratigraphic occurrence

Girvanella problematica occurs only in one sample (unit K-23), approximately 16.5 metres below the top of the Douro Formation in Section K (Appendix I). The masses of filaments are abundant in this sample. This is the only occurrence of *Girvanella* in the Douro Formation. Other occurrences may have been destroyed by the abundant burrowing organisms that generated the mottled dolomitic limestones. Garrett (1977) noted the destruction of thin mats of *Schizothrix* by surface bioturbation and by grazing organisms in Holocene sediments.

The matrix is a sparse biomicrite with zones of neomorphic sparry calcite. Bioclasts include coral, brachiopod, ostracod, echinioderm, bryozoan, gastropod and dasycladacean algal fragments. Whole fossils of all but the dasycladacean algae *Vermiporella* occur in this sample.

Generally, the bioclasts are angular with no evidence of abrasion.

The masses of *G. problematica* are unique in that they are not fragmented or disrupted as are the other bioclasts. Several reasons for this may be proposed: 1) *G. problematica* is autochthonous to the sediment whereas the other bioclasts are allochthonous, 2) *G. problematica* is more resistant to wave or current action than are the other bioclasts and 3) *G. problematica* inhabitated the sediment after deposition rather than during. The former two hypotheses are preferred over the latter. In the latter case, *G. problematica* may have assumed the habit of a boring endolithic algae rather than a binding or encrusting habit. The preservation of the filaments in sparite pockets is very similar to the growth habit of Holocene boring algae documented by Garrett (1977, p. 140, Fig. 79A).

The presence of the bioclasts and the nature of the bioclasts attest to some degree of agitation but the retention of a micrite matrix and the angularity of the bioclasts indicates predominantly quiet water conditions. As is also evident with *Girvanella* sp. in the oncoliths, *Girvanella* preferentially occurs in areas of at least intermittent agitation and show some resistance to abrasion by wave or current action. No sedimentological parameters occur laterally or stratigraphically adjacent to unit K-23.

The angular nature of the bioclasts indicates, that if indeed they are allochthonous, they are locally derived.

Regional stratigraphic trends and local paleontological evidence suggests that this stratum formed in a shallow subtidal environment, below or near wave base.

Girvanella sp.

(Plate 2, Figs. 3 and 4; Plate 9, Figs. 2 and 3)

Description

Thallus consists of small, nonseptate, tubular filaments with a well defined wall. Outside diameters range from 25-40 microns, being of uniform diameter within an individual mass. Filament diameter varies between masses on single bedding planes or in different units. Walls, which are 3-5 microns thick, are composed of microcrystalline calcite. The filaments are usually filled with anhedral calcite. No branching of the filaments occurs and there is no prefered orientation of the filaments relative to the growth surface.

Remarks

Girvanella sp. forms dense masses of filaments, unlike the loosely felted filaments of *Girvanella problematica*. The filaments are slightly larger (25-40 microns vs. 9-17 microns) in diameter than other Silurian *Girvanella* and the range of filament diameter between nodules indicates that several species of *Girvanella* may occur in the same stratum. However, the *Girvanella* found in the oncoliths will not be divided into species. To separate the filaments into

species, solely on the basis of filament diameter, would potentially introduce a number of species that would be relatively meaningless in the discussion of the paleoenvironment and the stratigraphic occurrence of the oncoliths.

Paleoenvironment and stratigraphic occurrence

There is some controversy over the validity of *Girvanella* as a paleodepth indicator. Most workers (Logan *et al.*, 1964; Wray, 1967a; Noble, 1970; Machielse, 1972; Lauritzen and Worsley, 1974; Peryt and Peryt, 1975; Gebelein, 1976; Flügel, 1977; Peryt, 1977; Toomey and LeMone, 1977; Tsien and Dricot, 1977) have considered *Girvanella* as a good facies indicator of relatively shallow marine environments such as drowned reef flats, submerged shoals, lagoons, shallow near shore areas, lagoons and reef complexes. Riding (1975), however, argued that *Girvanella* and other algae are not valid as depth indicators in paleoenvironmental studies.

Oncoliths usually suggest a low energy sedimentary regime with intermittent agitation by wave or bottom currents (Logan *et al.*, 1964; Noble, 1970; Bosellini and Ginsburg, 1971; Flügel, 1977; Milliman, 1977; Peryt, 1977). Both *Girvanella* sp. and *G. problematica* preferentially grew in environments that were subjected to some degree of agitation. Although oncoliths range to 100 metres of water depth they are most common in less than 5 metres of water

(Bosellini and Ginsburg, 1971; Buchbinder, 1977).

Two morphotypes, spherical and lobate, are recognized in the Upper Silurian strata from eastern Prince of Wales Island (Fig. 4.2a). The term *morphotype* refers specifically to final oncolith morphology even though oncoliths may change their external morphology during growth (Fig. 4.2b). Thus, an oncolith which was initially spherical may later develop a lobate morphology (Plate 10, Figs. 1 and 2). The morphotype of the oncoliths show no relationship to lithology and little relationship to the composition and shape of the nucleus.

Spherical morphotypes (Fig. 4.2a) are dominant, occurring as the only morphotype in nine of the sixteen oncolith occurrences. Lobate morphotypes (Fig. 4.2a) occur as the only morphotype in three occurrences, and together with spherical morphotypes in four occurrences (Table 4.1). Where the two morphotypes occur together there is no dominance of one over another.

Four oncolith types are defined on the basis of nucleus composition and the presence or absence of laminae. Type A oncoliths (Plate 2, Figs. 1 and 2) have a nucleus of the same composition as the surrounding stratum. This oncolith type is usually spherical and is laminated. They are rare and usually occur in association with other oncolith types but can occur as the only oncolith type (Table 4.2).

Type B oncoliths (Plate 9, Fig. 1) have a megafossil nucleus, usually a gastropod or a coral and lack laminae.

UNIT	SPHERICAL	LOBATE
R-59		X
R-61		X
R-62	X	
M-72	X	
M-73	X	
N-141	X	
H-37	X	
Q-38	X	X
Q-40	X	X
Q-41	X	
Q-42	X	
Q-44	X	X
Q-47	X	X
Q-48	X	
Q-50		X
Q-53	X	

TABLE 4.1. Occurrences of oncolith morphotypes.

They may be spherical or lobate; their morphotype being partly controlled by nucleus shape, since the *Girvanella* masses mimic it. This is the rarest oncolith type and occurs in unit Q-44 (Table 4.2).

Type C oncoliths are similar to Type B but are laminated. They may be spherical or lobate, and occur alone or with other oncolith types (Table 4.2).

Type D oncoliths (Plate 9, Fig. 4) have a nucleus comprised of a different lithology than the surrounding stratum. The oncolith has laminae and may be spherical or lobate. This is the most abundant oncolith type and can occur by itself or in association with other oncolith types (Table 4.2).

Oncolith bearing strata tend to occur in a stratigraphic sequence. The best example of an oncolith

UNIT	TYPE A	TYPE B	TYPE C	TYPE D
M-72				X
M-73	X			
N-141				X
H-37				X
R-59				X
R-61			X	X
R-62				X
Q-38				X
Q-40				X
Q-41				X
Q-42			X	X
Q-44		X	X	X
Q-47				X
Q-48	X		X	X
Q-50			X	
Q-53				X

TABLE 4.2. Occurrences of oncolith types.

succession occurs in Section PQ where nine of the sixteen oncolith occurrences span a 35 metre interval in the lower part of the Somerset Island Formation (Fig. 4.3). In Sections R (R-59, R-61, R-62) and M (M-72, M-73) the oncoliths also occur in sequential stratigraphic units. Only two of the sixteen oncolith occurrences (H-37, N-141) were isolated occurrences. This suggests that once conditions conducive to the formation of oncoliths were established, this state remained stable for a period of time.

Although two morphotypes and four oncolith types are recognized, they are purely descriptive terms and have little or no validity in paleoenvironmental interpretation. Bosellini and Ginsburg (1971) and Elliot (1975) suggested that oncoliths whose nuclei have the same composition as the surrounding sediment formed penecontemporaneously with that

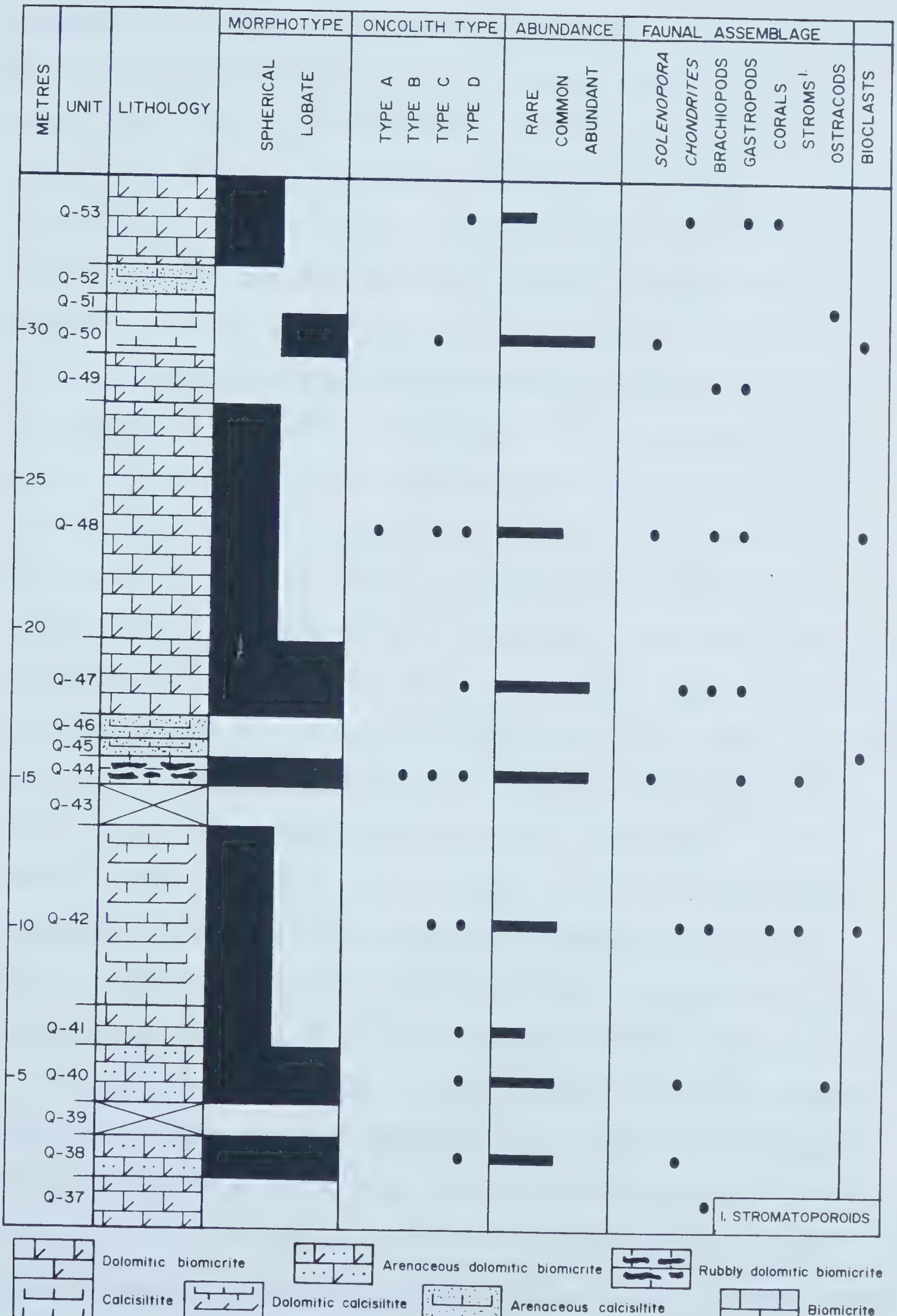


FIG. 4.3. Oncolith characteristics throughout a stratigraphic sequence, Section Q, lower Somerset Island Formation.

sediment. The presence of different nucleus types demonstrates only the availability and diversity of nucleation sites. A diversity of nucleus types were identified in Recent oncoliths in Shark Bay, West Australia (Logan *et al.*, 1964). Since all morphotypes and oncolith types are only descriptive terms and all morphotypes and nucleus types can occur together in a stratum, they will not be differentiated for paleoenvironmental interpretation.

Logan *et al.* (1964) considered the arrangement of laminae in oncoliths to be indicators of the frequency of agitation by wave or current action. Type "I" spheroidal structures (Fig. 4.2a) form in areas of low agitation, whereas Type "R" and Type "C" spheroidal structures (Fig. 4.2a) reflect the periodicity of agitation. Type "C" forms in areas of constant agitation while Type "R" forms in areas of intermittent agitation. Logan *et al.* (1964) observed all three spheroidal structures in the low intertidal to shallow subtidal zones in Shark Bay. Types "R" and "C" spheroidal structures equate with the spherical morphotypes of this study, whereas the lobate morphotypes may represent an extreme form of Type "I" spheroidal structures (Fig. 4.2a).

The problem with associating different energy regimes with morphotypes is that spherical and lobate morphotypes can occur together on a single bedding plane (Plate 7, Fig. 3). It can only be assumed that the oncoliths occurring together formed contemporaneously and were thus formed under the same sedimentary and energy regime. Oncoliths of both

morphotypes are in the same size range on any given bedding plane and therefore would have been subjected to the same hydraulic regime. The differences in morphotype and laminae arrangement may reflect the periodicity of movement of the oncolith not the strength of the agitation. Peryt (1977) considered oncolith size to be an indication of turbulence; the larger the oncolith, the stronger the turbulence.

It may be suggested that not all oncoliths were detached from the substrate at all times during growth. Lobate forms may reflect at least temporary attachment to the substrate, allowing growth to continue on only the exposed oncolith surface. The laminae of the lobes show similar doming and attachment along the margins as that of columnar stromatolites.

The strata bearing oncoliths usually contains a low diversity fauna of *Solenopora*, *Chondrites*, indeterminate small smooth shelled brachiopods, gastropods, solitary corals, tabular stromatoporoids and bioclastic fragments. The number of individuals of any taxon is also low. In only one unit (R-62) were the brachiopods (*Atrypoidea netserki*), found in association with the oncoliths, identifiable.

The fauna associated with the oncoliths indicates a marine environment; however, the low numbers of individuals and the low diversity suggests a more marginal marine environment than that of the underlying Douro Formation. The only occurrence of oncoliths in the Cape Storm Formation (H-37) is stratigraphically above a stromatolite bed which

probably formed in the low intertidal zone.

Eleven of the oncolith occurrences are in biomicrites, with the remaining five occurrences in calcisiltites or dolosiltites (Fig. 4.4). The latter usually contain a higher bioclastic fraction than the former. No relationship between morphotype and substrate lithology was observed. The association of oncoliths with micrite would suggest a low energy environment (Noble, 1970). No diagnostic sedimentological or paleontological criteria occurs in association with the oncoliths of any stratum.

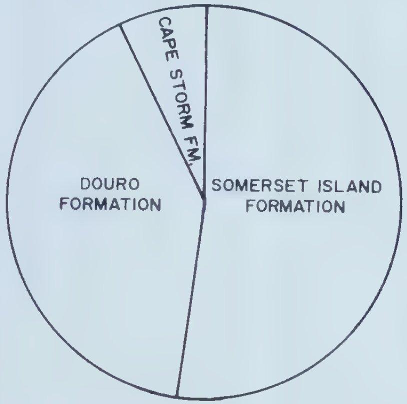
Sedimentological evidence suggests that the strata containing the oncoliths formed in areas with slow sedimentation rates. Lithoclasts of the same composition as the surrounding stratum suggest contemporaneous sedimentation, lithification (cementation) and disruption of the sediment. The sediment must have been partially lithified for intraclasts to form, but not enough to provide a stable substrate for stromatolites to form. Oncoliths preferentially form on soft, loose substrates, whereas stromatolites preferentially form on hard, stable substrates (Gebelein, 1976; Monty and Mas, 1981).

Erosion surfaces and/or planar hardgrounds (Kennedy and Garrison, 1975) are present in the oncolith beds (Plate 10, Fig. 2) suggesting minor hiatuses in sedimentation. Fürsich (1979) recognized that the lack of a mineralized surface in association with hardgrounds suggests that it was either a short term hiatus or that erosion was important in these

DISTRIBUTION OF RHODOLITHS
(*Solenopora* nodules)

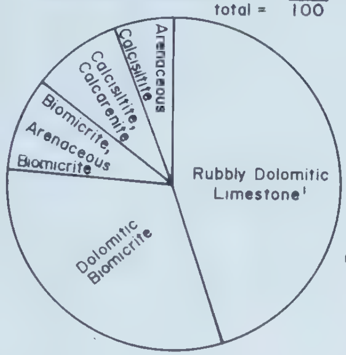
BY: FORMATION

	%	N
CAPE STORM FORMATION	7	9
DOURO FORMATION	41	51
SOMERSET ISLAND FORMATION	52	64
total =	100	124



BY: LITHOLOGY

	%	N
Rubbly Dolomitic Limestone[1]	45	56
Dolomitic Biomicrite	31	38
Biomicrite, Arenaceous Biomcrite	9(7,2)	12 (9,3)
Calcisiltite, Calcarenite	9(6,3)	12 (8,4)
Arenaceous Calcisiltite	6	6
total =	100	124

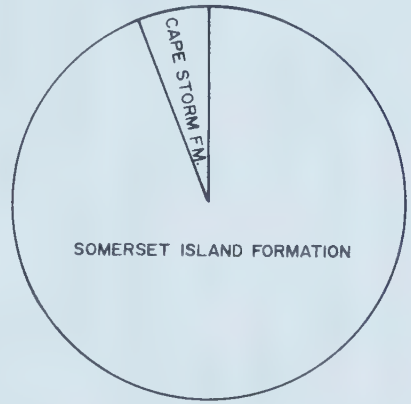


1. includes:
Rubbly Dolomitic Biomicrite
Rubbly Dolomitic Calcarenite

DISTRIBUTION OF ONCOLITHS
(*Girvanella* nodules)

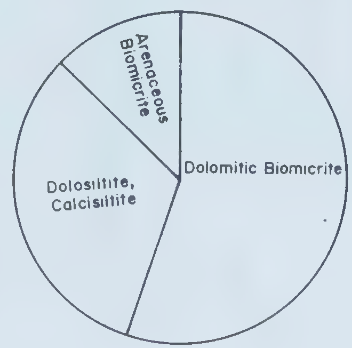
BY: FORMATION

	%	N
CAPE STORM FORMATION	6	1
DOURO FORMATION	—	—
SOMERSET ISLAND FORMATION	94	15
total =	100	16



BY: LITHOLOGY

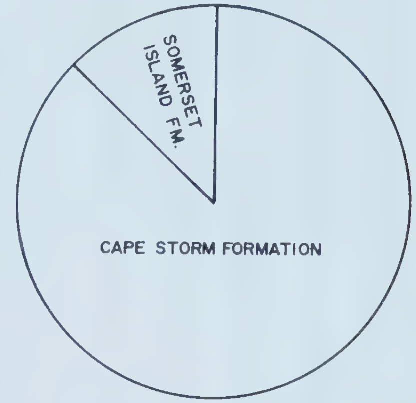
	%	N
Dolomitic Biomicrite	55	9
Dolosiltite, Calcisiltite	32 (19,13)	5(3,2)
Arenaceous Biomicrite	13	2
total =	100	16



DISTRIBUTION OF STROMATOLITES

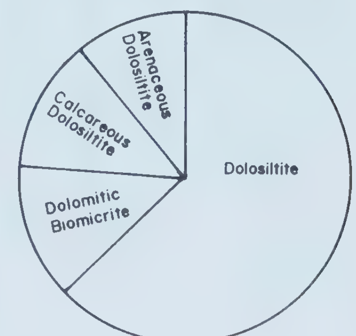
BY: FORMATION

	%	N
CAPE STORM FORMATION	90	27
DOURO FORMATION	—	—
SOMERSET ISLAND FORMATION	10	3
total =	100	30



BY: LITHOLOGY

	%	N
Dolosiltite	63	19
Dolomitic Biomicrite	13	4
Calcareous Dolosiltite	13	4
Arenaceous Dolosiltite	11	3
total =	100	30



N = Number of occurrences

FIG. 4.4. Distribution of algae in Upper Silurian strata on eastern Prince of Wales Island.

hardgrounds. Both these factors were possibly important in the development of these erosion surfaces.

Due to the lack of any diagnostic criteria in the oncolith beds or in proximal stratigraphic units, paleoenvironmental interpretation is based primarily on regional stratigraphic trends. The dominant occurrence of oncoliths is in the lower part of the Somerset Island Formation. In most sections, the lower part of the Somerset Island Formation contains desiccation polygons, stromatolites, intraclast breccias, ripple marks and cross-laminated calcisiltites and calcarenites which are indicative of shallow subtidal to low intertidal environments. However, the strata of the Somerset Island Formation which contain the oncoliths do not show evidence of desiccation and were probably deposited in the shallow subtidal zone above wave base.

The oncoliths show little to no abrasion of the outer laminae and are considered autochthonous to the strata in which they occur. Other megafossils from the oncolith beds show little evidence of transportation and are also considered autochthonous to the strata in which they occur.

The fauna found in association with the oncoliths suggests a marginal marine environment with no implication of paleodepth. Formation of oncoliths requires some agitation but not necessarily turbulent conditions. Most of the oncoliths from eastern Prince of Wales Island are spherical or were spherical at some time during their

growth, suggesting formation at or above the wave base where there was relatively constant agitation. The oncoliths from eastern Prince of Wales Island possibly formed in shallow, near-shore subtidal environments with low rates of sedimentation.

Spongiostromata

Stromatolites

The study of stromatolites is essentially one of morphogenesis and is reflected by the three schools of thought regarding stromatolite classification: 1) morphology reflects time dependent biological factors, 2) morphology reflects environmental conditions and 3) morphology reflects an interaction of abiotic and biotic factors. No classification scheme is universal in its application but instead appears to to be unique to the fossil suite of a particular study and the author's interpretation of morphogenesis. Hofmann (1969) presented an excellent review of stromatolite classification schemes and it will not be reiterated here.

Logan *et al*. (1964, p. 69) defined stromatolites as laminated structures composed of particulate sand, silt and clay sized sediment which have been formed by the trapping and binding of detrital sediment particles by an algal film. The emphasis on lamination is important; Kalkowski (1908) and Semikhatov *et al*. (1979) regarded laminations (stromatoids) as the fundamental building block of

stromatolites. Semikhatov *et al.* (1979) considered each lamination as a couplet composed of two sub-laminae, the algal binding layer and the attached sediment layer. The stromatolites in the Upper Silurian strata on eastern Prince of Wales Island are distinctly laminated and composed of silt sized and smaller carbonate grains.

The stromatolite classification scheme (Table 4.3) used in this study is modified from Donaldson (1963). This simple classification scheme, based on external morphology, is preferred over the schemes using Linnean nomenclature based on form taxa or the more complicated morphological classification schemes based on external morphology, shape of laminae, linkage and stacking. Three stromatolite types of Donaldson (1963) occur in the Upper Silurian strata on eastern Prince of Wales Island (Table 4.3).

The stromatolites in the Upper Silurian strata on eastern Prince of Wales Island are accurately described by the classification scheme of Donaldson (1963). Only minor revision of the classification of the digitate types is required. Thus, the following terminology of Donaldson (1963, p. 7) is used:

1) "Hemispherical stromatolites (*Collenia* Walcott, 1914): consist of successive hemispherical, convex-upward laminations. Laminations are uniform thickness and roughly concentric with respect to the centre of the basal surface."

2) "Digitate stromatolites (*Gymnosolen* Steinmann, 1911): consist of discrete cylindrical columns which

FORM CLASSIFICATION Donaldson (1963)	FORMS PRESENT (In this study)
Hemispherical (*Collenia* Walcott, 1914)	Hemispherical
Bulbous (*Cryptozoon* Hall, 1883)	
Columnar (*Archaeozoon* Matthew, 1890)	
Undulatory (*Weedia* Walcott, 1914)	Undulatory
Digitate (*Gymnosolen* Steinmann, 1911)	Digitate
Pisolitic (*Pycnostroma* Gurich, 1906)	

TABLE 4.3. Stromatolite types (from Donaldson, 1963).

commonly branch upwards into parallel columns of lesser diameter."

Digitate stromatolites (*Tungussia* Semikhatov, 1962)(this study): consist of discrete cylindrical columns which coalesce laterally and vertically and are of variable diameter. These digitate stromatolites differ from those of Donaldson (1963) in the style and nature of the branching.

3) "Undulatory stromatolites (*Weedia* Walcott), 1914: consist of laterally continuous laminations which have irregular wavy boundaries."

Paleoenvironment and stratigraphic occurrence

The Cape Storm Formation contains the most abundant and diverse stromatolite types found in the Upper Silurian

strata on eastern Prince of Wales Island. Twenty-seven of the 30 stromatolite occurrences are in the Cape Storm Formation (Fig. 4.4). Twenty-six of these 27 occurrences are hemispherical stromatolites (Plate 7, Fig. 4) found in association with undulatory stromatolites (Plate 11, Figs. 2 and 3). Digitate stromatolites occur only in unit L-29 (Plate 12, Figs. 1 and 2). The other three occurrences are hemispherical stromatolites found in the lower part of the Somerset Island Formation. These stromatolites are similar in external morphology to those in the Cape Storm Formation but are not always associated with undulatory stromatolites. No stromatolites occur in the Douro Formation. Since the hemispherical stromatolites of the Cape Storm and Somerset Island formations are similar in size, shape, structure and associated paleontological and sedimentological parameters; they are initially discussed as a single group.

Differential weathering highlights the occurrence of stromatolites in the field since the stromatolites are more resistant than the surrounding rock (Plate 7, Figs. 1, 2 and 4). Individual stromatolites are typically 30-50 cm in diameter and 20-40 cm in amplitude, being elliptical in cross-section parallel to depositional strike. When exposure permitted observation, the individual stromatolites appear to be elongate perpendicular to depositional strike. Maximum size of an individual stromatolite is 1.6 m in diameter and 1.1 m in amplitude. The stromatolites can occur as isolated hemispheres, as clusters on bedding planes or as stacked

hemispheres (Plate 7, Fig. 4).

Lateral spacing of the stromatolites is usually greater than the diameter of the stromatolites, ranging from 0.65 m to 3.2 m. Measurement of the lateral spacing was restricted to the two dimensions along depositional strike; the third dimension perpendicular to depositional strike was not exposed.

Locally, stromatolites form small bioherms. In unit B-26, stacked stromatolites form a bioherm 8.5 m in width, measured parallel to structural strike, and 2.1 m high. The third dimension was unobtainable due to poor exposure. Individual stromatolites are 50.0 cm in diameter and 25.0 cm in amplitude. The base of the bioherm is laterally continuous with a unit containing stromatolites and the peripheries of the bioherm intertongue with dolosiltites of laterally adjacent strata.

The base of attachment of the hemispherical stromatolites appears as a convex-upward, disc-shaped hollow in which the laminae extend beyond the hollow to attach to the substrate. Although this may suggest some uniform mechanism of initiation of hemispherical growth, there are some differences in the stromatolites of the Cape Storm and Somerset Island formations. Doming of Recent stromatolites is attributed to four factors (Logan *et al.*, 1964): 1) lateral expansion of the algal mat, 2) doming over pre-existing irregularities, 3) more active sediment binding on the highs and inhibition of the mats in the lows and 4)

evolution of gases beneath the mats.

The hemispherical stromatolites grade laterally and vertically into undulatory stromatolites and algal laminites. This is most evident in the Cape Storm Formation in which undulatory stromatolites, hemispherical stromatolites and algal laminites are common. In these strata, initiation of doming was possibly due to the lateral expansion of the algal mats. Maintenance and accentuation of the hemispherical stromatolites may possibly have been by scour along the margins or differences in sedimentation rates between the interareas and the highs.

The dolosiltites of the interareas onlap the hemispherical stromatolites and show a downwarping and thickening in the centre of the interareas. These dolosiltites are commonly rippled; the ripple marks are symmetrical, aligned parallel to strike and have sinuous crestlines. Amplitude of the ripple marks is less than 1 cm and the wavelength is 11-15 cm. Locally, synaeresis cracks occur on the rippled dolosiltites.

The formation and maintenance of hemispherical stromatolites is dependent on the inability of the active algal mats to transgress the interareas or the destruction of the mat in the interareas (Logan *et al.*, 1964, p. 79). This statement is supported by features found in association with the hemispherical stromatolites in the Cape Storm Formation: 1) the rippled dolosiltites in the interareas indicate a mobile substrate inhibiting lateral growth of the

algal mat, 2) the intraclasts of the algal laminites (Plate 11, Fig. 4) and of the dolosiltites suggest scour and destruction of the algal mats in the interareas and 3) the increase in bed thickness from the margins of the stromatolites into the interareas suggests that burial rates may have exceeded the growth rate of the algal mat in the interarea.

The hemispherical stromatolites of the Somerset Island Formation are not always associated with undulatory stromatolites and algal laminites as are those in the Cape Storm Formation. Doming may be related to irregularities in the basal platform (Plate 7, Fig. 2) rather than lateral expansion of the algal mat. The basal platforms of the hemispherical stromatolites of the Somerset Island Formation are more irregular and have a maximum relief of 4.0 cm. Doming, however, does not appear to be restricted to the highs on the basal platform. In the Cape Storm Formation, the algal structures occur in a sequence, whereas in the Somerset Island Formation the stromatolites are usually found as single beds in Sections D and M at Strzelecki Harbour.

Stromatolites in the Upper Silurian strata on eastern Prince of Wales Island usually occur in dolosiltites and rarely in micrites; only 4 of the 30 stromatolite occurrences are in micrites (Fig. 4.4). This is in contrast to the oncoliths in which 11 of 16 occurrences are in micrites and 5 occurrences are in calcisiltites or

dolosiltites. The oncoliths in Sections Q and R occur in the same stratigraphic interval in the Somerset Island Formation as the hemispherical stromatolites in Sections D and M. This difference in lithology supports the proposition that oncoliths preferentially formed on unstable, soft substrates whereas the stromatolites preferentially formed on stable, hard substrates.

The difference in lithology may indicate a lower degree of turbulence during deposition of the micrites which contain the oncoliths than the dolosiltites which contain the stromatolites. Gebelein (1976) noted that Recent oncoliths formed in similar environments to stromatolites but were subjected to periodic turbulence. The differences in substrate lithology between the oncoliths and the stromatolites suggests the strong influence of abiotic controls on morphogenesis. It should, however, be restated that the stromatolites are a spongiostromate algal structure and that the oncoliths are a porostromate algal structure. The differences in morphogenesis may, in part, be biotic controlled and reflect the preferential selection of the algae to specific environments. Both the oncoliths and the stromatolites of this study appear to have been sensitive to clastic detritus; only 2 of 16 occurrences and 3 of 30 occurrences, respectively, occur in association with quartzose carbonates (Fig. 4.4).

In unit L-22, three dimensional exposure of the stromatolites showed them to be elongate perdendicular to

depositional strike. Logan (1961), Hoffman (1967), Gebelein (1969), Logan *et al.* (1974) and Gebelein (1976) noted a similar asymmetry in Recent and Ancient stromatolites; the stromatolites being elongate parallel to the direction of wave propagation or current action and normal to the shoreface. These authors also documented thickening of the laminae in the direction of sediment supply. No thickening of the laminae is apparent in the stromatolites of this study but it would appear that the elongation developed normal to depositional strike. The dominant current direction, during the formation of the stromatolites, must have been in an east-west direction with the sediment supply predominantly from the Boothia Horst to the east.

The hemispherical stromatolites of the Cape Storm and Somerset Island formations are associated with a low diversity fauna of ostracods and high spired gastropods. In the Cape Storm Formation, the gastropods locally form packstone lenses in the interareas and adjacent stata. The limited fauna of ostracods and gastropods would suggest a slightly restricted environment but not the hypersaline conditions associated with Recent stromatolites in Shark Bay, West Australia by Logan *et al.* (1964). Little evidence was found during this study to indicate hypersaline conditions during formation of the stromatolites.

The hemispherical stromatolites of the Cape Storm Formation commonly have desiccation polygons on their crests while the interarea strata contain rippled dolosiltites and

intraclasts. The crests of the hemispherical stromatolites must have been at least intermittently exposed, whereas the interarea strata reflect continual wetting or limited subaerial exposure without desiccation. These stromatolites possibly formed in the low intertidal zone.

Atrypoidea erebus occurs in the basal platform of the stromatolites in unit M-45 in the lower part of the Somerset Island Formation. The association of brachiopods with the stromatolites of the lower part of the Somerset Island Formation indicates a more open marine environment but proximal to the low intertidal zone as indicated by desiccation features in strata above and below the stromatolite beds.

The digitate stromatolites (Plate 12, Figs. 1 to 4) in unit L-29 occur in the middle of the Cape Storm Formation, approximately 2.5 m above a hemispherical stromatolite bed. The associated fauna consists of high spired gastropods and *Planolites*, the gastropods forming packstone lenses. Serial sectioning of the digitate stromatolites, showed the digits to be variable in diameter but forming discrete columns that can coalesce laterally and vertically. The nature of the intersection of the laminae and the margins of the digits (Plate 12, Fig. 3) indicates that there has been some destruction of the digits along the outer margins. The common association of gastropods with the digitate stromatolites may suggest biological destruction of what may have originally been columnar stromatolites. The destruction

of Holocene algal mats by grazing and burrowing organisms was documented by Garrett (1970, 1977).

There is no evidence of desiccation in unit L-29 or in stratigraphically adjacent units (Appendix I). Brachiopods occur 1.4 m above and 1.7 m below this unit and the presence of omission surfaces 4.9 m above unit L-29 indicate subaqueous rather than subaerial exposure. The presence of hemispherical stromatolites 2.5 m below the digitate stromatolites indicates a low intertidal environment. The digitate stromatolites were possibly formed in a low intertidal to shallow subtidal environment.

Thrombolites

Thrombolites (Aitken, 1967) (=pseudostromata of Wolf, 1965a; algal ooze of Machielse, 1972) are characterized by a clotted appearance, lack of laminae and a microfabric consisting of micrite and calcite cement (Wolf, 1965a; Aitken, 1967; Machielse, 1972). The only occurrence of thrombolites (unit A-66) in the Upper Silurian strata on eastern Prince of Wales Island is approximately 10.5 metres below the top of the Douro Formation in Section A (Appendix I). In unit A-66, four growth phases of algae are present (Plate 13, Figs. 1 and 2); growth phases 1 to 3 are similar in morphology and form the columns of the thrombolite. Growth phase 4 is the final phase of algal development, forming discrete caps on the columns and extending laterally onto the interarea sediments.

The relief expressed by the top of phase 4 possibly represents original depositional topography (Plate 13, Figs. 1 and 2). Hofmann (1969, p. 36) suggested that the laminae in stromatolites represent microbathymetry during the time interval at which each laminae was at the active interface between the bound sediment below and the moving water with suspended particles above.

Immediately overlying the thrombolites are laminated, quartzose dolosiltites which grade upwards into argillaceous micrites (Plate 13, Fig. 3). Algal laminites occur at the base of the unit. Sedimentary structures occurring in unit A-66 include symmetrical ripple marks with an amplitude of less than 1 cm and a wavelength of 10-15 cm, desiccation polygons and intraclasts. No megafossils occur in this unit but stratigraphically adjacent units contain ostracods, solitary rugose corals and abundant brachiopods.

Aitken (1967) postulated a hypersaline environment for thrombolite development, whereas Machielse (1972) recognized algal ooze predominantly in slightly restricted subtidal environments and extending into the intertidal zone. The thrombolites in unit A-66 possibly formed in a relatively low energy, very shallow subtidal to low intertidal environment as indicated by the associated sedimentary structures. The low diversity fauna would suggest a slightly restricted environment or a stressed geobiocoenose but not hypersaline conditions.

Fenestrae and algal laminites

The term *fenestrae* is restricted to lenticular voids in a laminar or nonlaminar configuration in carbonate rocks that are presumed to have been formed in association with blue-green algae (Machielse, 1972, p. 193). Algal laminites refer to thinly-laminated carbonate rocks, presumed to have been formed in association with blue-green algae but did not develop into distinct structures that can be classified as stromatolites.

Fenestrae occur in all three formations of the Upper Silurian strata on eastern Prince of Wales Island, but predominantly in the Cape Storm Formation. The fenestrae are usually aligned subparallel to the laminae (Plate 14, Fig. 3), occurring either alone or in association with algal laminites. Typically, the fenestrae are partially filled or floored by a fine calcisiltite, the remaining void space being filled by sparry calcite (Plate 14, Fig. 4).

In the Cape Storm and Somerset Island formations, the fenestrae occur in association with algal laminites composed of calcisiltites and dolosiltites. The only occurrence of fenestrae in the Douro Formation (unit N-35) is unique in that the fenestrae occur in a micrite and are not found in association with algal laminites. Unit N-35 is anomalous relative to stratigraphically adjacent units in that the only megafossil component are rare echinoderm fragments. Large-scale ripple marks occur at the upper boundary of the unit. The ripple marks are oblique to depositional strike

and have straight, parallel crestlines, with a maximum amplitude of 4.2 cm and a maximum wavelength of 22.0 cm (Plate 15, Fig. 1). Underlying the unit are 2.3 metres of skeletal calcarenite and overlying the unit is a 0.8 metre coral-stromatoporoid biostrome.

Unit N-35 was probably deposited on a shallow subtidal environment such on the crest of a submerged shoal. The skeletal calcarenite in unit N-34 was possibly deposited as a shallow submerged shoal. The presence of a skeletal calcarenite and the ripple marks suggests relatively constant agitation by wave or current action. Similar large-scale ripple marks on skeletal calcarenite banks in the Lower Muschelkalk of southwestern Germany were documented by Schwarz (1975).

Logan *et al.* (1974, p. 147) documented algal mats and associated fenestral fabrics from shallow water depths of 4 m below low water level to about 2 m above low water level in the supratidal zone. Shinn (1968a) recognized fenestral fabrics as being indicative of the supratidal zone, whereas Machielse (1972) observed fenestral fabrics in very shallow subtidal to supratidal zones, interpreted as lagoonal or protected environments of reef complexes and carbonate shoals. Fenestrae in the Upper Silurian stata on eastern Prince of Wales Island predominantly occur in strata that was probably deposited in the intertidal zone and rarely in the very shallow subtidal zone.

Algal laminites occur in the Cape Storm Formation, rarely in the Douro Formation and as thin beds in the Somerset Island Formation. Stratigraphic distribution of the algal laminites and the fenestrae are similar but algal laminites are more common. Fenestrae from the Upper Silurian strata on eastern Prince of Wales Island occur, with only one exception (unit N-35), in association with algal laminites, whereas the algal laminites do not always contain fenestrae. Typically, the algal laminites occur in calcisiltites or dolosiltites.

The algal laminites generally weather recessive and rarely was it possible to determine the algal laminite (mat) type. The crinkled algal laminite in unit Q-67 (Plate 14, Fig. 1) is similar to the colloform mat of Logan *et al.* (1974, p. 147, Fig. 4); however, the crinkling is approximately an order of magnitude smaller. In cross-sectional view, the laminae are planar and show a poorly developed fenestral fabric; the fenestrae are aligned subparallel to the laminae. Logan *et al.* (1974, p. 150) observed colloform mats in the subtidal zone from low water level to depths of about 4 metres. Associated fauna and sedimentary structures in unit Q-67 suggest the unit was deposited in a shallow subtidal environment at or near low tide level.

Most of the algal laminites of this study are planar-laminated (Plate 14, Fig. 2) and coalesce laterally with hemispherical stromatolites. Undulatory stromatolites

occur as lateral and vertical transition units between planar algal laminites and hemispherical stromatolites (Plate 11, Figs. 1, 2 and 3). The algal laminites are usually unfossiliferous and contain desiccation polygons, intraclasts and brecciation of the laminites (Plate 11, Fig. 4). Algal laminites usually occur in strata deposited in the low intertidal zone and less commonly in the subtidal zone.

C. Chlorophycophyta

Dasycladaceae

Phylum CHLOROPYHCOPOHYTA Papenfuss, 1946

(green algae)

Family DASYCLADACEAE Kuetzing, 1843,

ortho mut Stizenberger, 1860

Genus VERMIPORELLA Stolley, 1893

Type species

Vermiporella fragilis Stolley (1893, Plate 8, Figs. 7-11).

Diagnosis

Thallus is an irregular, bent or sinuous cylinder of varying diameter. Walls are penetrated by circular to polygonal shaped pores (branches or rays) that are disposed perpendicular to the axis of the central stem. Pores are

numerous, closely set, of uniform diameter and do not bifurcate into a secondary series. The fossils are usually fragmentary and poorly preserved.

Discussion

This genus is poorly documented in the literature which, in part, may reflect the poor quality of preservation of the fossils. Rezak (1959, p. 120) and Johnson (1961, p. 125) considered *Vermiporella* to be a primitive and irregular dasycladacean algae. Riding (1975, p. 205), however, argued that *Vermiporella* is of a foraminiferal affinity and should be excluded from the dasyclads.

Vermiporella? sp.

(Plate 8, Figs. 1 and 2)

Description

Thallus are cylindrical masses. Outside diameters of the calcareous body range from 357 to 470 microns. Tangential and oblique sections show the crude polygonal shape of the pores. Pores are numerous, 20-24 microns in diameter and uniform in diameter. Transverse sections, showing the nature of the branching, are absent.

Remarks

Vermiporella sp. occurs as poorly preserved fragments in a biomicrite. The specimens are tentatively assigned to this genus due to the similarity in character to

Vermiporella described by Johnson and Konishi (1959), Johnson (1961) and Wray (1967a). The fragments show some similarity to another Silurian dasyclad *Rhabdoporella* Stolley of the same tribe. The genus *Vermiporella* is retained with the algae and not reassigned to the foraminifera.

Paleoenvironment and stratigraphic occurrence

Vermiporella sp. occurs in only one sample from unit K-23, approximately 16.5 metres below the top of the Douro Formation in Section K (Appendix I). The fragments occur together with abundant *Girvanella problematica* masses and the discussion of the paleoenvironment of this genus is included with the *G. problematica*.

D. Rhodophycophyta

Solenoporaceae

Phylum RHODOPHYCOPHYTA Papenfuss, 1946

(red algae)

Family SOLENOPORACEAE Pia, 1927

Genus SOLENOPORA Dybowski, 1877

Type species

Solenopora spongioides Dybowski (1877, Plate 2, Figs. 11a and 11b).

Diagnosis

Thallus consists of nodular masses, usually hemispherical or spherical forms and rarely encrusting forms. In transverse section, the cells are polygonal to rounded. In vertical section, the cell threads are radially or vertically arranged from the base of the thallus. Cross partitions separating the cells in the threads are irregularily spaced or absent, usually thinner and less conspicuous than the vertical wall of the threads.

Discussion

Genera of Paleozoic Solenoporaceae are distinguished on the basis of the character of their internal cellular tissue (Table 4.4). Pia (1927) recognized three genera (*Solenopora*,

Pseudochaetetes and *Parachaetetes*), whereas Johnson (1961) and Wray (1967a; 1977) recognized only two genera (*Solenopora* and *Parachaetetes*) (Table 4.4). The classification scheme of Johnson (1961) and Wray (1967a; 1977) appears to be more common in the literature and is used in this study.

Brooke (1984), however, argued that *Solenopora* shows only a superficial similarity to the calcareous algae and is in "fact a collection of unrelated organisms and is probably polyphletic." One of the species under discussion is *S. filiformis* which Brooke (1984) removed from this genus but did not assign it to any other. This species occurs in the rhodoliths of this study and is retained with the *Solenopora*.

Solenopora sp.

(Plate 4, Fig. 4; Plate 16, Figs. 1 and 2)

Description

Thallus consists of irregular shaped masses with a maximum dimension of 12.0 cm. In transverse section, the cells are hexagonal and vary in diameter, ranging from 90-230 microns. Cell walls are 3-6 microns thick and no cell length was obtained. In vertical section, cell threads are vertical to radially arranged, without partitions.

PALEOZOIC GENERA	CHARACTERISTIC FEATURES
Solenopora	Vertical walls of cell threads dominant. Cross partitions of the threads are usually thin, less conspicuously developed than vertical walls and irregularly spaced.
Pseudochaetetes	Cross partitions of cell threads are common, thick and irregularly spaced.
Parachaetetes	Cross partitions separating the cells are well defined, thick and regularly spaced. The tissue has a grid-like appearance in vertical section.

TABLE 4.4. Characteristic features of Paleozoic Solenoporaceae.

Remarks

The cell size of *Solenopora* sp. is larger than that reported for other Paleozoic solenopores except that of *S. gippslandicum* Chapman, 1920 from the Middle Devonian of Victoria, Australia. *S. gippslandicum* occurs as subspherical nodules with a maximum individual cell diameter of 150 microns. Most cell diameters of Paleozoic *Solenopora* are less than 60 microns and more commonly between 25-50 microns (Johnson, 1960, p. 21, Table 1). The wide range of cell sizes may indicate more than one species of *Solenopora* forms these bioherms.

Paleoenvironment and stratigraphic occurrence

Solenopora in the Upper Silurian strata on eastern Prince of Wales Island usually occur as rock-contributing constituents such as rhodoliths rather than as a framework constituents. In the Somerset Island Formation (units Q-55, Q-61, Q-63, Q-71, N-73 and G-104), *Solenopora* sp. occurs as algal bioherms. The size and shape of the colonies are unknown due to the nature of the rock; the bioherms occur in either rubbly dolomitic biomicrites or fractured, planar-bedded biomicrites. Surface weathering as well as the rubbly weathering of these rock types mask the geometric limits of the algal bioherms.

Maximum size of the fragmented algal lumps is 12.0 cm but the irregular outline of these lumps and the proximity of these lumps to one another suggests that they once formed a mound of coalescing algal colonies. Filling between the algal lumps, composed of *Solenopora* sp., are dolomitic biomicrites which show a gradational contact with the algae. The nature of this contact appears to show destruction of the algae in favour of micrite (Plate 16, Figs. 1 and 2; Plate 18, Fig. 2). Wolf (1965b) termed this destruction of algae, grain diminution of algae to micrite.

The *Solenopora* bioherms probably formed in shallow subtidal environments with some degree of water agitation. The associated fauna consists of brachiopods and ostracods. Ripple marks are commonly associated with the algal bioherms.

Solenopora filiformis Nicholson, 1888

(Plate 17, Figs. 1, 2 and 4)

Description

Thallus consists of nodular masses usually as subspherical forms but rarely as encrusting forms. In transverse section, the individual cells are hexagonal to rounded. Cell diameter is variable, ranging from 25-40 microns. No cell length was measured. The cell walls are 3-5 microns thick. In vertical section, cell threads are radially arranged, without cross partitions or very irregularily spaced cross partitions. Tissue is compact.

Remarks

S. filiformis is the dominant species of *Solenopora* in the Upper Silurian strata on eastern Prince of Wales Island. The taxonomy of this algae was studied using transverse and vertical sections of some nodules and vertical sections of other nodules. Recognition to the generic level is possible in the field by etching freshly fractured nodules and on weathered surfaces of nodules (Plate 17, Fig. 3). Spheroidal nodules are dominant but mammalated forms also occur. A single encrusting form was found in unit B-59. Mean nodule size is 1.5 cm. and maximum size is 9.0 cm. *S. filiformis* is not restricted in its stratigraphic distribution and occurs with a variety of lithologies and faunal associations.

Paleoenvironment and stratigraphic occurrence

Rhodoliths (=rhodolites, =rhodoids) are nodular growth forms comprised primarily of coralline (red) algae (Bosellini and Ginsburg, 1971; Heckel, 1975; Belka, 1979; Bosence, 1983a; 1983b; Burgess and Anderson, 1983) that grew as individual entities contemporaneously with the deposition of the sediment but detached from the substrate (Bosellini and Ginsburg, 1971; Bosence, 1983a). Bosellini and Ginsburg (1971) and Bosence (1983b) attempted morphological classifications of rhodoliths on the basis of nodule sphericity relative to the long, short and intermediate axis of the nodules. Bosence (1983b) applied this method to Recent and Ancient rhodoliths and argued that morphology is a useful indicator of hydraulic energy, water depth, geography and temperature. Much of this work was based on branching density of Recent rhodoliths comprised of *Lithothamnium corallioides*.

No branching rhodoliths occur in the Upper Silurian strata on eastern Prince of Wales Island. Two morphologies of rhodoliths occur; spheroidal forms which range from spherical to ellipsoidal, and mammalated forms which have short rounded protuberances on the surface of the rhodolith (Plate 17, Fig. 1). Mammalated rhodoliths equate with the columnar rhodoliths of Bosence and Pedley (1982) and Bosence (1983b). The term *mammalated* is restricted to rhodoliths and the term *lobate* is restricted to oncoliths This distinction is based on internal structure, in rhodoliths the

protuberances are in continuous growth with the thallus, whereas in oncoliths the protuberances are independent growths on the surface of the oncolith.

Mammalated rhodoliths occur in eight stratigraphic units and always in association with spheroidal rhodoliths (Table 4.5). The mammalated rhodoliths are generally larger (3.0 cm vs. 1.5 cm) than the spheroidal rhodoliths from the same stratum. Rhodoliths occur with oncoliths in six stratigraphic units, three occurrences are with lobate oncoliths, two are with spherical oncoliths and one is with both oncolith morphotypes (Table 4.5). Lobate oncoliths and mammalated rhodoliths do not occur together in the same stratum.

As postulated for the development of lobate oncoliths, the protuberances on rhodoliths may represent a restriction on the movement of the rhodolith such as temporary attachment to the substrate. Some of the rhodoliths show fragmentation on only one side and the nature of this edge suggests a torn attachment from the substrate rather than abrasion and transportation. Mammalated rhodoliths show little evidence of abrasion and are possibly autochthonous to the stratum rather than allochthonous. This contradicts the argument of Bosence and Pedley (1982) that columnar rhodoliths were formed by the abrasion and transportation of branching rhodoliths.

Although lobate oncoliths and mammalated rhodoliths both result from at least temporary attachment to the

UNIT	RHODOLITHS		ONCOLITHS	
	MAMMALATED	SPHEROIDAL	LOBATE	SPHERICAL
M-55	X	X		
M-56	X	X		
Q-78	X	X		
Q-80	X	X		
Q-82	X	X		
N-88	X	X		
N-90	X	X		
N-107	X	X		
Q-44		X	X	X
Q-48		X		X
Q-50		X	X	
R-59		X	X	
R-62		X		X
R-61		X	X	
M-72				X
M-73				X
N-141				X
H-37				X
Q-38			X	X
Q-40			X	
Q-41				X
Q-42				X
Q-47			X	X
Q-53			X	

TABLE 4.5. Association of rhodolith forms and/or oncolith morphotypes.

substrate, the differences in substrate type should be noted. Oncoliths, whether lobate or spherical, preferentially occur in micrites and presumably soft substrates. The mammalated rhodoliths preferentially occur in calcisiltites or calcarenites and presumably more stable substrates. This agrees with the observation of Bosellini and Ginsburg (1971) that Recent columnar rhodoliths occur only on stabilized substrates such as sand.

In unit Q-50, lobate oncoliths occur with spheroidal rhodoliths in a calcisiltite matrix (Fig. 4.3, Table 4.5). If the factors controlling morphology were solely abiotic, mammalated rhodoliths would be expected to occur in association with this substrate type. As the lobate oncoliths and the mammalated rhodoliths appear to be mutually exclusive of another, even in favourable environments, it would suggest that there is some biotic control on morphology.

Rhodoliths from the Upper Silurian strata on eastern Prince of Wales Island occur with three faunal and lithological associations: 1) a diverse faunal assemblage in which the megafossils and the rhodoliths are fragmented and abraded. The matrix is usually a rubbly dolomitic limestone in which the limestone lumps are composed of skeletal calcarenite (Plate 18, Fig. 3), 2) a diverse faunal assemblage in which the megafossils and rhodoliths show little to no abrasion or disruption. The matrix is usually a rubbly dolomitic limestone in which the limestone lumps are

composed of biomicrite and 3) a less diverse faunal assemblage of gastropods, brachiopods, ostracods and *Chondrites* or *Planolites*. The matrix is usually a planar bedded biomicrite, calcisiltite or calcarenite.

Twelve occurrences of rhodoliths are with the high faunal diversity rubbly dolomitic calcarenites, ten of these occurrences are in the Douro Formation and two are in the Somerset Island Formation. The calcarenites are bioclastic in origin with echinoderm fragments being the most common grain type. Megafossils and/or rudite sized allochems include; *Solenopora*, *Megalomoidea*, echinoderms, stromatoporoids, corals (*Favosites* dominant), gastropods and brachiopods. The allochems are usually rounded and have pronounced micrite envelopes (Bathurst, 1971). Megafossils are generally fragmented and may show several cycles of fragmentation and encrustation by stromatoporoids and *Favosites*.

The association of *Solenopora* and *Megalomoidea* predominantly occurs in the rubbly dolomitic calcarenites. The *Megalomoidea* valves tend to be disarticulated and aligned parallel to bedding, usually convex upwards. The orientation of the valves suggests that the valves were subjected to some wave or current action during deposition. *Solenopora* occurs as abraded nodules or rarely as encrusting masses on *Megalomoidea* valves. The abraded nodules may be allochthonous to the calcarenites.

The rubbly dolomitic calcarenites were deposited in relatively high energy environments such as submerged shoals and were subjected to constant wave or current action. The rounded skeletal grains, alignment of the elongate bioclasts and void spaces filled by a single phase syntaxial calcite cement attest to this. Locally, pockets of micrite and microspar occur in association with a finer grained calcarenite or grainstone.

The rubbly nature of the dolomitic calcarenites masks any sedimentary structures, if indeed any are present. In only two stratigraphic units proximal to the calcarenites are there any sedimentary structures indicative of wave action. Unit N-35, a micrite between a calcarenite and a biostrome, contains large ripple marks up to 4.2 cm in amplitude and 22.0 cm in wavelength (Plate 15, Fig. 1). In Section PQ, a calcareous quartz arenite (units Q-19, Q-20) with large scale cross-laminations occurs aproximately 6 metres above a sequence of rubbly weathering calcarenites (Appendix I).

Elliot (1975), Heckel (1975) and Orszag-Sperber *et al.* (1979) documented rhodoliths in abraded, skeletal calcarenites which contain a diverse faunal assemblage of echinoderms, formainiferas, bryozoans, pelecypods, gastrods, corals, dasycladacean algal fragments and intraclasts. The association of rhodoliths with a calcarenite suggests a more open marine environment such as a shoal or back-reef area (Heckel, 1975; Tsien and Dricot, 1977, Belka, 1979).

Narbonne (1981) and Narbonne and Dixon (1982) recognized solenoporacean algae in the initial Crinoid Stage of lithistid sponge reefs in the Douro Formation on Somerset Island. The Crinoid Stage is a stabilization stage during which mounds of skeletal debris, dominated by crinoids, bryozoans and solenoporacean algae, are stabilized and accumulated (Narbonne, 1981; Narbonne and Dixon, 1984). The height of these mounds can be a few decimetres (Narbonne,1981; Narbonne and Dixon, 1984). This stage of the reefal mounds is similar to the shoals which occur on Prince of Wales Island; however, no reefal development succeeded these shoals.

The second association of rhodoliths with a high diversity faunal assemblage and a rubbly dolomitic biomicrite predominantly occurs in the Douro Formation and rarely in the Cape Storm or Somerset Island formations. The megafossil assemblage is similar to that of the rubbly dolomitic calcarenites but can also include trilobites and othoconic nautiloids. The megafossils are intact and show no signs of abrasion; neither the rhodoliths nor the bioclasts show evidence of abrasion or transportation. The rhodoliths are spheroidal, low in population numbers and small, rarely exceeding 1.5 cm in diameter. These units were probably deposited in a relatively deeper, subtidal environment below or near normal wave base.

The association of rhodoliths with a low diversity faunal assemblage of gastropods, brachiopods, ostracods and

Chondrites or *Planolites* occur in planar-bedded, variably quartzose biomicrites, calcisiltites and calcarenites of the Cape Storm and Somerset Island formations. Unlike the skeletal calcarenites, these calcisiltites and calcarenites are composed of calcite grains of indeterminate origin. The rhodoliths are more abundant than in either high diversity faunal assemblage. Locally, the rhodoliths can be very abundant and form rhodolith pavements (Plate 17, Fig. 1; Plate 18, Fig. 1). Both spheroidal and mammalated rhodolith forms occur in these calcisiltites and calcarenites. Macrofossils and rhodoliths may be intact or fragmented but show little evidence of abrasion.

Two sedimentological situations occur with this third type of rhodolith association: 1) strata which contain desiccation polygons, intraclasts, cross-laminations and ripple marks and 2) stata void of sedimentary structures. The former is the least common and may represent single events such as storm deposits on tidal flats in which the rhodoliths may be allochthonous. Abate *et al.* (1977) noted a similar sedimentological and paleontological association with rhodoliths in storm deposits on tidal flats in the Triassic of northwestern Sicily. The strata of the Upper Silurian on eastern Prince of Wales Island which contain rhodoliths but lack sedimentary structures were probably deposited in relatively shallow water with little agitation by wave or current action.

Rhodoliths preferentially occur in carbonate environments free of clastic detritus. Only 9 of the 124 occurrences of rhodoliths in the Upper Silurian strata on eastern Prince of Wales Island occur in association with clastic detritus (Fig. 4.4). Thirty occurrences of rhodoliths are in calcisiltites and calcarenites, the remaining 94 occurrences are in biomicrites. Milliman (1974) and Heckel (1975) noted that rhodoliths are sensitive to terrigenous detritus and preferentially form in clear water carbonate regimes. Belka (1979), however, argued that rhodoliths occur in association with carbonate muds containing detrital quartz grains. The diversity and stratigraphic range of the rhodoliths in the Upper Silurian strata on eastern Prince of Wales Island show that no single environment is unique to rhodolith development.

E. Summary

The algal taxonomy, used in this study, appears to have resolved the problems of taxonomy, in particular the problems of Porostromata vs. Spongiostromata and *Sphaerocodium* vs. *Wetheredella*. There appears to be no overlap of "families" or of genera using this modified nomenclature. The only remaining question is that of the affinity of *Wetheredella*. Until more work is done on this genus and further occurrences are found, this affinity may have to remain unresolved.

Morphogenesis, at least in this algal suite, appears to be a function of abiotic rather than biotic factors. The external morphologies of oncoliths and rhodoliths and the association of the respective morphotypes and forms (Table 4.5) suggests abiotic factors control morphology. While all four morphologies do not occur together, there is sufficient overlap between and within nodule morphologies to indicate environmental factors dominate over biotic factors. Biotic factors can not be totally excluded but in the case of rhodoliths in which both forms are comprised of the same species, biotic factors may be minimal relative to abiotic factors.

Biotic factors appear to be significant between the species of *Solenopora*. No overlap of growth forms was observed between the nodular growth forms of *S. filiformis* and the biohermal growth habit of *Solenopora* sp. Both species appear to have markedly different growth forms.

In the Somerset Island Formation, stromatolites occur in the same stratigraphic interval as oncoliths, but in different localities. This indicates a widespread geographical distribution of conditions conducive to the formation of algal structures. Stromatolites preferentially occur in dolosiltites whereas the oncoliths preferentially occur in micrites. This suggests that environmental factors controlling the sediment type ultimately controlled the type of algal structure. The laminations in the lobes of the oncoliths are similar to the laminae structure in the

hemispherical stromatolites, merging with the margins of the lobe and becoming convex-upward in the lobe.

Of note, however, is the dominance of stromatolites on the transgressive tidal flats and of oncoliths on the regressive tidal flats. Oncoliths occur in only one stratum during the transgressive phase and stromatolites occur in only three strata during the regressive phase. The respective environments conducive to the formation of algal structures are replicated during both phases of the transgressive-regressive cycle but the distribution of the algae is not. The mechanism responsible for this distribution is unknown.

The algal complex in unit M-56 shows a hemispherical shape on the bedding plane. The individual columns comprised of *Girvanella* sp., *Sphaerocodium* sp. and *Wetheredella* sp. form domes similar in shape but smaller than the hemispherical stromatolites of this study. The similarity between the structures of the algal complexes, the lobes of the oncoliths and of the hemispherical stromatolites further supports the argument for environmental factors vs. biological factors controlling morphology. Regardless of the algal type, it is possible to generate similar external morphologies.

The other spongiostromate structures (thrombolites, algal laminites and fenestrae) do not have a porostromate equilavent. Their distribution may reflect the distribution of environments conducive to that particular algae,

suggesting both abiotic and biotic factors on the formation of these algal structures.

Although a diversity of algal structures and associations are found in the Upper Silurian strata on eastern Prince of Wales Island; few, if any, are reliable paleoenvironmental indicators if not found in association with other paleontological or sedimentological criteria. Generally, it may be stated that all the algal structures occur in the intertidal to shallow subtidal zone with little environmental distinction beyond that. Depth criteria relied on the presence or absence of desiccation features and indicators of wave or current action. The argument of Riding (1975) that algae are not useful as paleodepth indicators appears to have some merit in the strata of this study. The only environmental parameter that algae may be useful for is the relative turbulence as reflected in the lithology, condition of the algal structure and internal structure of the algae such as laminae.

V. NATURE OF THE MOTTLED DOLOMITIC LIMESTONES

A. Introduction

Mottled dolomitic limestones constitute a large proportion of the Upper Silurian strata on eastern Prince of Wales Island, occurring in the upper member of the Cape Storm Formation, the Douro Formation and the lower part of the Somerset Island Formation. These mottled dolomitic limestones form a relatively continuous sequence approximately 250 metres thick.

These rocks are composed of two mutually exclusive lithologies, limestone and dolostone, that show virtually no intermixing. Based on the distribution of these two component lithologies, a classification scheme for the dolomitic limestones was developed (this study; Jones *et al.*, 1979). It is the physical nature and the problem of genesis of these rocks that commands interest in them.

A succession of depositional, biogenetic and diagenetic processes acted concomitantly to form the mottled dolomitic limestones. While each process is relatively simple in its own merit, the temporal and spatial interactions of these processes form a complicated synthesis. This synthesis relies heavily on the law of uniformitarianism and the work of Bromley (1967, 1968, 1975), Kennedy (1967) and Kennedy and Garrison (1975) on the Cretaceous Chalks of southern England.

The mottled dolomitic limestones of the Upper Silurian strata on eastern Prince of Wales Island are reasonably homogenous in their heterogeneity, suggesting an uniformity and continuity of process. Jones *et al.* (1979) addressed the issue of genesis of the rubbly dolomitic and argillaceous limestones of the Read Bay Formation on Somerset Island and suggested a similar genesis as proposed by this study but hedged on the role of bioturbation in the formation of these rocks. Bioturbation is acknowledged as having played a major role in the genesis of these mottled dolomitic limestones.

The first section of this chapter is purely descriptive and is included for the purposes of clarification of terms leading into the discussion of the genesis. A taxonomic discussion of the ichnogenera is presented because of the role that these burrowing organisms had in the formation of these rocks.

The second section is a chronological synthesis of the processes responsible for the formation of the mottled dolomitic limestones and the interactions of these processes. The rates of bioturbation, sedimentation and lithification are acknowledged as being the dominant controls on the formation of mottled dolomitic limestones and their variations. A cyclicity of processes is evident when these processes are examined in a temporal context.

B. Description of mottled dolomitic limestones

The Upper Silurian dolomitic limestones contain two lithologic components; limestone lumps or layers (Jones, 1974; Jones and Dixon, 1977, Jones *et al.*, 1979) and a dolosiltite matrix. Both components occur as discrete and mutually exclusive lithologies. The term *dolosiltite* is used as a descriptive field term; it refers to the size of the dolomite grains, not their genesis.

Three distinct types of mottled dolomitic limestone are delineated on the basis of the distribution of the limestone lumps or layers relative to the dolosiltite matrix (Fig. 5.1). The ratio of limestone to dolosiltite is highly variable both within and between units; the dolosiltite ranges from being partings between semi-continuous limestone lumps or layers to totally isolating the limestone lumps. No mean ratio can be estimated but generally limestone constitutes a greater percentage of the rock; rarely do the dolosiltites form more than 50% of the rock.

Types I and II rubbly mottled dolomitic limestone have been described by Jones *et al.* (1979, p. 232, Table 1). In Type I, the limestone forms discrete lumps which generally have no preferred orientation relative to bedding (Plate 19, Fig. 5). In Type II, the limestone forms semi-continuous layers aligned subparallel to bedding (Plate 19, Fig. 5). The rubbly argillaceous limestones of Jones *et al.* (1979) do not occur in the Upper Silurian strata on eastern Prince of Wales Island.

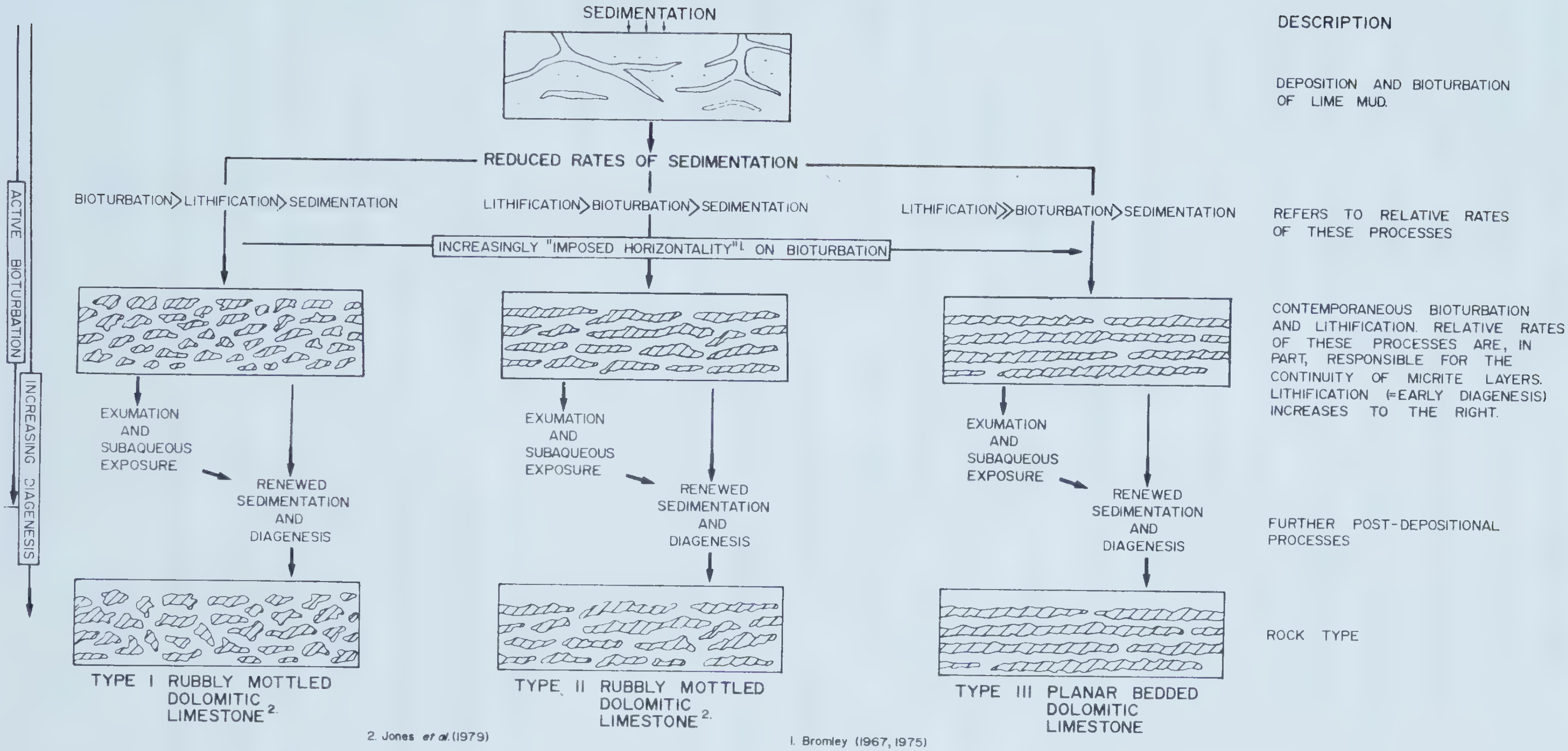


FIG. 5.1. Schematic diagram showing the genesis of dolomitic limestone types in the Upper Silurian strata on eastern Prince of Wales Island (for full explanation, see text).

A third type, Type III, of mottled dolomitic limestone is recognized in the strata of this study (Plate 15, Fig. 4). In this type, the limestone forms relatively continuous layers with irregular upper and lower surfaces and has the appearance of being interbedded with the dolosiltite matrix. Not all the interbedded limestones and dolostones are classified as Type III; only those which are similar in appearance to the Types I and II are assigned to this type. Other interbedded limestones and dolostones show sedimentary structures that suggest a different origin than the mottled dolomitic limestones. These lithologies also differ in appearance from the mottled dolomitic limestones.

Types I and II have a pronounced rubbly weathering (Plate 20, Figs. 1 and 3) whereas Type III tends to weather more blocky or planar. The limestone weathers prominant relative to the surrounding dolosiltite matrix and the rubbly weathering reflects the continuity of the limestone lumps or layers.

The three types of mottled dolomitic limestone are attributed to a similar genesis, their differences caused by local variations in the rates of the controlling processes (Fig 5.1). Types I and II have a closer genetic link than does Type III to either of the above types. The former types occur as lateral and vertical equivalents both within and between units whereas Type III occurs only in vertical succession with Types I and II.

Types I and II are most common, forming greater than 75% of the Douro Formation and occurring in the lower part of the Somerset Island Formation. Type III occurs in the upper member of the Cape Storm Formation, the Douro Formation and the lower part of the Somerset Island Formation.

Petrography of the limestone lumps/layers

The limestone lumps or layers of the mottled dolomitic limestones are predominantly sparse biomicrites but range from fossiliferous micrites to skeletal calcarenites. Approximately one third of the limestones are peloidal, containing greater than 10% peloids (Plate 15, Fig. 3). The limestones are typically medium brown-grey on a fresh surface, weathering medium grey. Bioclasts consist of brachiopod, coral, bryozoan, gastropod, coralline algae, echinoderm, bivalve, trilobite, ostracod and stromatoporoid fragments. Complete body fossils of most of the above taxa also occur.

The limestone lumps or layers commonly contain low amplitude stylolites and small fractures filled by sparry calcite. The amplitude of the stylolites is generally less than 5 mm and the stylolites are orientated horizontally or obliquely to bedding. The fractures are usually orientated oblique to bedding and terminate at the dolosiltite-limestone interface, rarely extending into the dolosiltite matrix. Where the fractures and stylolites

intersect, the fractures are offset across the stylolite. Stylolites can terminate at the dolosiltite-limestone interface or be laterally continuous with microstylolite seams.

Petrography of the dolosiltite matrix

The dolosiltite matrix is generally light grey to buff on fresh surfaces but ranges from yellowish-orange to greenish-grey. Weathered surfaces are typically buff in color but can show the same range of colors as fresh surfaces. Locally, malachite occurs in the dolosiltite matrix and the greenish-grey color of the matrix is attributed to the influence of copper-rich solutions.

The dolosiltite matrix (Plate 21, Fig. 3) is composed of euhedral dolomite rhombs in a microcrystalline calcite matrix (idiomorphic texture of Beales, 1953; idiotopic texture of Friedman, 1965; Morrow, 1978; idiotopic-E of Gregg and Sibley, 1984). Mean rhomb size is 40 microns, a measurement based on the mean rhomb size in approximately 235 samples. In a single sample or between samples the dolomite rhombs are relatively equigranular. Maximum rhomb size is 120 microns.

Typically, dolomite rhombs constitute 50-65% of the matrix. Calcite in the form of micrite, microspar or bioclasts forms the other 35-50%. Locally, calcite occurs as a replacement mineral in the dolomite rhombs (dedolomitization). With increasing dolomitization, the

euhedral rhombs give way to subhedral and less commonly anhedral dolomite grains (hypidiotopic and xenotopic textures of Friedman, 1965; Morrow, 1978; idiotopic-S of Gregg and Sibley, 1984). Accessory constituents of the dolosiltite matrix include azurite, malachite, hematite, pyrite, quartz and rare muscovite grains. Quartz occurs as detrital grains or as a replacement mineral in the dolomite rhombs. Jones *et al.* (1979), using x-ray diffraction analysis, found the noncarbonate portion of the rubbly dolomitic limestones of the Read Bay Formation on Somerset Island to be commonly less than 10% of the whole rock.

The bioclastic content of the dolosiltite matrix is less than that of the limestone lumps or layers but is of a similar diversity. Complete body fossils are rare in the dolosiltite matrix and are not dolomitized. Similar petrographic relationships were noted by Morrow (1978) in the Upper Ordovician mottled dolomitic limestones on Devon Island.

The dolosiltite-limestone interface

The interface of the two lithologies is generally sharp and highly irregular (Plate 21, Figs. 1, 2 and 4; Plate 22, Figs. 1, 2 and 3). Neither the dolomite rhombs of the matrix nor the bioclasts of either the limestone or the matrix transgress the interface. In rare situations, a thin trace of opaques lines the interface (Plate 23, Fig. 3) and is laterally continuous with flaser structures (Kennedy and

Garrison, 1975 ; Garrison and Kennedy, 1977) or microstylolite seams (Wanless, 1979). These opaques may occur along the top or bottom interface of the limestone lumps or layers but are most common along the bottom interface (Plate 23, Fig. 3).

When the interface is gradational (Plate 21, Fig. 3), there is a zone extending for several hundred microns in which euhedral dolomite rhombs occur in the micrite of the limestone lumps or layers (porphrotopic of Friedman, 1965; idiotopic-P of Gregg and Sibley,1984) (Plate 21, Fig. 3). In hand sample or in the field, the interfaces appear sharp (Plate 22, Figs 1, 2 and 3) and it is only in thin section that the gradational contact is apparent. A gradational interface is most common where the dolosiltite matrix forms isolated lenses rather than a continuous dolosiltite layer. The micrite may also grade into a microspar towards the interface. Morrow (1978, p. 297, Fig. 6) observed a similar "halo of microspar" adjacent to the burrows in the mottled dolomitic limestones of the Upper Ordovician strata on Devon Island.

In all three types of mottled dolomitic limestone, the dolosiltite-limestone interface is similar. On bedding plane views and in three dimensions, the dolosiltite matrix is irregular in distribution but the patches of matrix are connected (Plate 19, Fig. 4). When viewed in a polished slab, cut perpendicular to bedding, the patches of matrix do not always appear to be connected but may show communication

along a zone that can be best described as a stylolite swarm or parting (Plate 23, Fig. 3). Along these partings, the dolosiltite matrix is poorly developed. Braun and Friedman (1969) noted a similar situation in the mottled dolomitic limestones of the Lower Ordovician Tribe Hills Formation in which dolomitization locally followed stylolites.

Some of the limestone lumps in the Type I mottled dolomitic limestones have straight, vertical interfaces. Laterally adjacent angular lumps have a matching interface suggesting there has been post-lithification fracturing of the lumps. The angularity of these lumps differs from the characteristic rounded, irregular shapes of the typical lump.

C. Ichnotaxa

A description of the inchnotaxa present in the mottled dolomitic limestones is introduced since bioturbation played a major role in the formation of the mottled dolomitic limestones of the Upper Silurian strata on eastern Prince of Wales Island. Many workers (Beales, 1953; Osmond, 1956; Lebauer, 1965; Nichols, 1966; Bromley, 1967, 1968, 1975; Hallam, 1967; Kennedy, 1967, 1970, 1975; Matter, 1967; Thomas, 1968; Braun and Friedman, 1969; Fürsich, 1973; Goldring and Kazmierczak, 1974; Kennedy and Juignet, 1974; Kennedy and Garrison, 1975; Kendall, 1977; Miller, 1977; Rubin and Friedman, 1977; Morrow, 1978; Jones *et al.*, 1979; Mullins *et al.*, 1979; Abed and Schneider, 1980; Narbonne,

1984; Pickerill *et al.*, 1984; Sheehan and Schiefelbein, 1984) attribute, at least in part, a biogenetic origin for mottled dolomitic or nodular limestones. Most of these workers considered bioturbation to be just one mechanism acting in concert with depositional and diagenetic controls to form mottled dolomitic or nodular limestones.

Bioturbation occurs in association with all three types of mottled dolomitic limestone. Three ichnogenera, *Thalassinoides*, *Chondrites* and *Planolites*, dominate. Kennedy (1975) considered such an assemblage to be representative of the *Cruziana* assemblage of Seilacher (1967). The ichnogenus, *Palaeophycus*, also occurs in association with the mottled dolomitic limestones but is less common. Since the mode and degree of preservation is highly variable, recognition of the burrow structures is often difficult. Commonly the burrows occur as sedimentary mottling and only rarely can the pattern and form of the burrow system be delineated.

Systematic Ichnology

Ichnogenus THALASSINOIDES

Ehrenberg, 1944

Thalassinoides sp.

(Plate 19, Figs. 1 and 2; Plate 25, Fig. 4; Fig. 5.2)

Description. Cylindrical burrows 1-3 cm in diameter and of a relatively constant diameter. Burrows have a Y- or

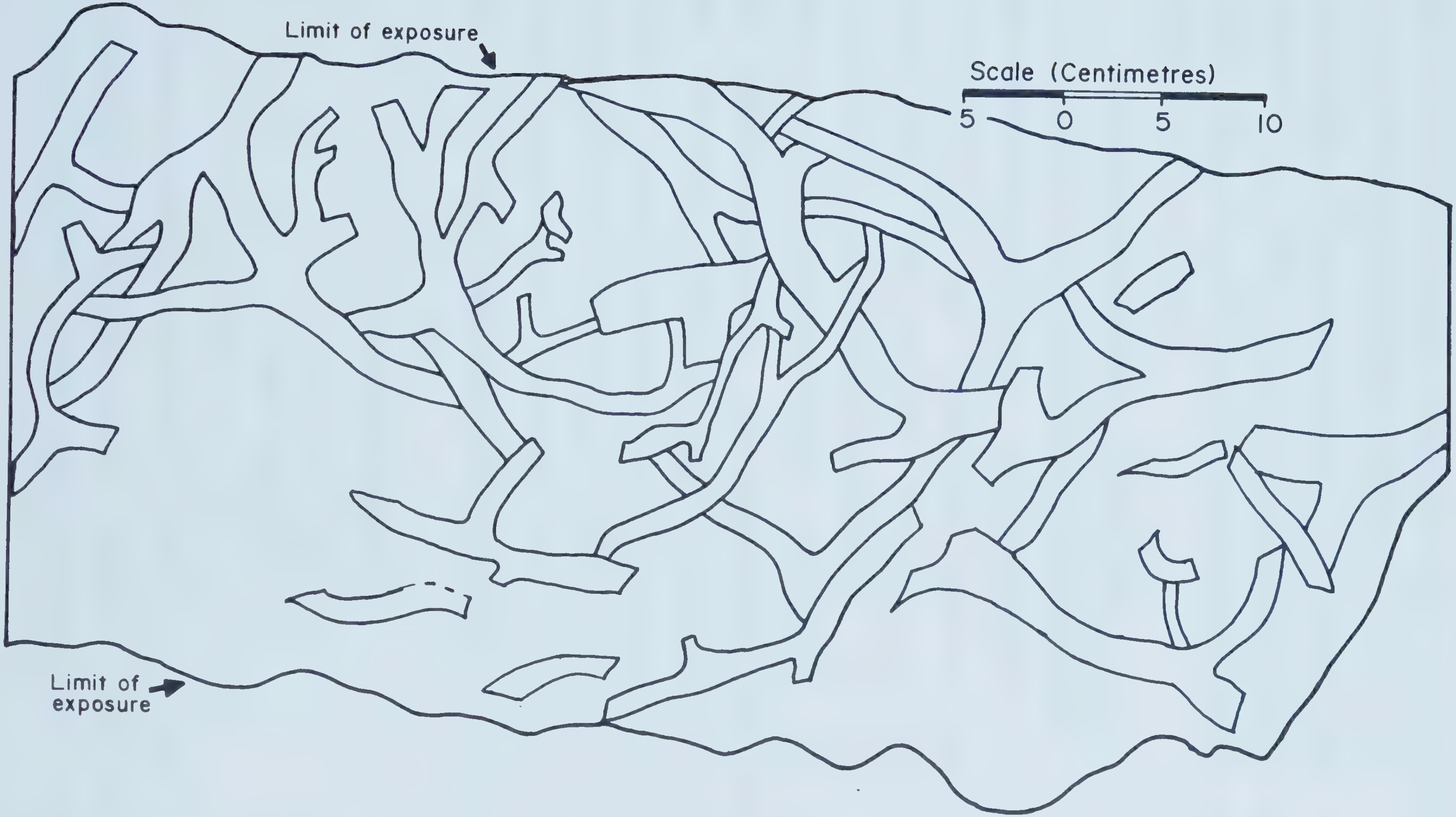


FIG. 5.2. Line drawing of *Thalassinoides*-type burrows exposed on a bedding plane in the Douro Formation, field unit L-59.

T-shaped branching habit and may or may not show inflation at the points of branching. Branching interval is approximately 5 cm; becoming highly irregular where the burrows are crowded. Horizontal systems dominate and may form a polygonal pattern on bedding planes. Burrow wall is smooth.

Remarks and occurrence. These burrow systems are similar to the burrows assigned to the ichnogenus *Thalassinoides* which ranges from the Ordovician to the Holocene (Bromley, 1967, 1968, 1975; Kennedy, 1967, 1975; Shinn, 1986b; Weimer and Hoyt, 1971, Fürsich, 1973; Kennedy and Juignet, 1974; Kennedy and Garrison, 1975; Miller, 1977; Abed and Schneider, 1980; Archer, 1984; Bromley and Ekdale, 1984; Ekdale and Bromley, 1984, Sheehan and Schiefelbein, 1984). *Thalassinoides* occurs only in association with the mottled dolomitic limestones and are filled with micrite, dolosiltite, or chert. Although dolosiltite burrow-fill is the most common mode of preservation, a micrite burrow-fill exposed in convex hyporelief shows the best preservation.

Ichnogenus PLANOLITES

Nicholson, 1873

Planolites sp.

(Plate 19, Fig. 3)

Description. Cylindrical to elliptical burrows usually 1 cm in diameter. The burrows are isodiametric, horizontal to slightly inclined, meander randomly and rarely branch.

Burrow wall is smooth.

Remarks and occurrence. Planolites occur only in association with the mottled dolomitic limestones. Preservation is generally poor and the burrows are usualy filled by dolosiltite and rarely by micrite. *Planolites* are less common than *Thalassinoides* or *Chondrites.*

Ichnogenus CHONDRITES

von Sternberg, 1833

Chondrites sp.

(Plate 26, Figs. 2, 3 and 4; Plate 27, Figs. 1 and 4)

Description. Cylindrical to elliptical burrows ranging from 1-3 mm in diameter. Vertical and horizontal systems are present but horizontal or slightly inclined systems dominate. Horizontal systems have a radial tendency and are dendritic, having greater than two orders of branching. Burrow wall is smooth.

Remarks and ocurrence. Chondrites is the most abundant ichnogenus, occurring throughout the Upper Silurian strata on eastern Prince of Wales Island except in the lower member of the Cape Storm Formation. These traces are not restricted to any single lithology. Preservation ranges from heterolithic mottling in which the burrows are filled by a different lithology than the surrounding stratum to monolithologic color mottling. The diversity of form in the *Chondrites* burrow systems may reflect several species or the control of substrate on the morphology of the burrow

systems.

Ichnogenus PALAEOPHYCUS

Hall, 1847

Palaeophycus sp.

(Plate 25, Fig. 2)

Description. Burrows are approximately cylindrical in cross-section, ranging in diameter from 1-3 cm. Branching habit is Y- or T-shaped with no inflation at the point of branching. Burrows are straight to slightly curved and have smooth walls. Only horizontal systems occur.

Remarks and occurrence. As with most of the ichnogenera previously described, *Palaeophycus* occurs only in association with the mottled dolomitic limestones, in particular Types I and II. The burrows are usually filled with micrite and are only apparent when exposed in convex hyporelief. *Paleophycus* are rare in the mottled dolomitic limestones.

Ichnogenus PHYCODES

Richter, 1850

Phycodes sp.

(Plate 26, Fig.1)

Description. Cylindrical burrows, ranging from 1.0-1.5 mm in diameter and forming horizontal bundled structures. Structures are 6-8 cm in length and circumscribe an arc of approximately 45°. The burrows have a smooth wall and may

show a single order of branching.

Remarks and occurrence. Phycodes are the least common burrow in the mottled dolomitic limestones, occurring in only Type III of the Somerset Island Formation. The burrows are filled with dolosiltite and are exposed in convex hyporelief. *Phycodes* is not considered part of the ichnocoenose responsible for generating the mottled dolomitic limestones.

Ichnogenus TRYPANITES

Mägdefrau, 1932

Trypanites sp.

Description. Simple unbranched, cylindrical borings with a single opening to the surface. The borings are 1.0-1.5 mm in diameter and 6-8 mm in length.

Remarks and occurrences. Trypanites are rare and occur as single borings in *Solenopora* nodules of the Douro Formation. The rhodoliths are commonly inverted with the boring orientated downwards and have a dark rim on the surface. This suggests that the nodules may be allochthonous to the stratum or at least locally transported and were subaqueously exposed as a hard substrate or in association with a hardground.

Preservation of ichnogenera

Although bioturbation is postulated as a major factor in the formation of the mottled dolomitic limestones of this study, the burrow systems are not always evident. Commonly, the burrow systems must be implied by the pattern of the sedimentary mottling and the presence of well preserved burrow systems in adjacent similar rock types. The lack of well developed burrow systems is attributed to the failure of preservation rather than the absence of burrow structures.

Pickerill *et al.* (1984) described a sequence of rubbly limestones from the Upper Ordovician in which the recognition of burrows became increasingly more difficult as the intensity of bioturbation increased. In the more extreme case of mottled units, recognition became impossible. Fürsich (1973) also noted that with an increase in burrow intensity, the individual burrows become indiscernible. These situations are analogous to the problem of trace fossil identification in the mottled dolomitic limestones of this study.

Biogenetic sedimentary mottling is recognized on at least two levels in the mottled dolomitic limestones. The most apparent are burrows in which the burrow-fill is of a different lithology than the surrounding stratum. The other is color mottling in the original sediment or burrow-fill. This is most commonly associated with *Chondrites* (Plate 19, Fig. 6; Plate 25, Figs. 1 and 3; Plate 26, Figs. 2 and 3)

but can also occur in the larger traces (Plate 22, Fig. 2; Plate 23, Fig. 3).

D. Genesis of mottled dolomitic limestones

The genesis of mottled dolomitic limestones in the Upper Silurian strata on eastern Prince of Wales Island is a difficult if not problematic issue; the very nature of the rocks masks the clues as to their origin. Prior to this study, Jones *et al*. (1979) were the only workers to address the problem of the genesis of the Upper Silurian mottled limestones found throughout the Arctic Archipelago.

The genesis of the mottled dolomitic limestones is attributed to a number of processes (Table 5.1) that act concomitantly and/or in chronological succession. Any synthesis proposed for the genesis of these rock types must consider these processes and address the observable characteristics of the mottled dolomitic limestones. Thus, it must explain: 1) the mutually exclusive nature of the two lithologic components, 2) the spatial distribution of these two components and the resultant three types of mottled dolomitic limestone (Fig. 5.1) and 3) the thick, continuous sequence of mottled dolomitic limestones. An index to the spatial and temporal relationships of these processes may be best described by examining the relationships of the burrow structures and the limestone lumps or layers.

The proposed synthesis is simple in process but difficult in demonstration. It will be assumed that in order

1. Depositional controls.
    a) Nature of original sediment.
    b) Rate of sedimentation.

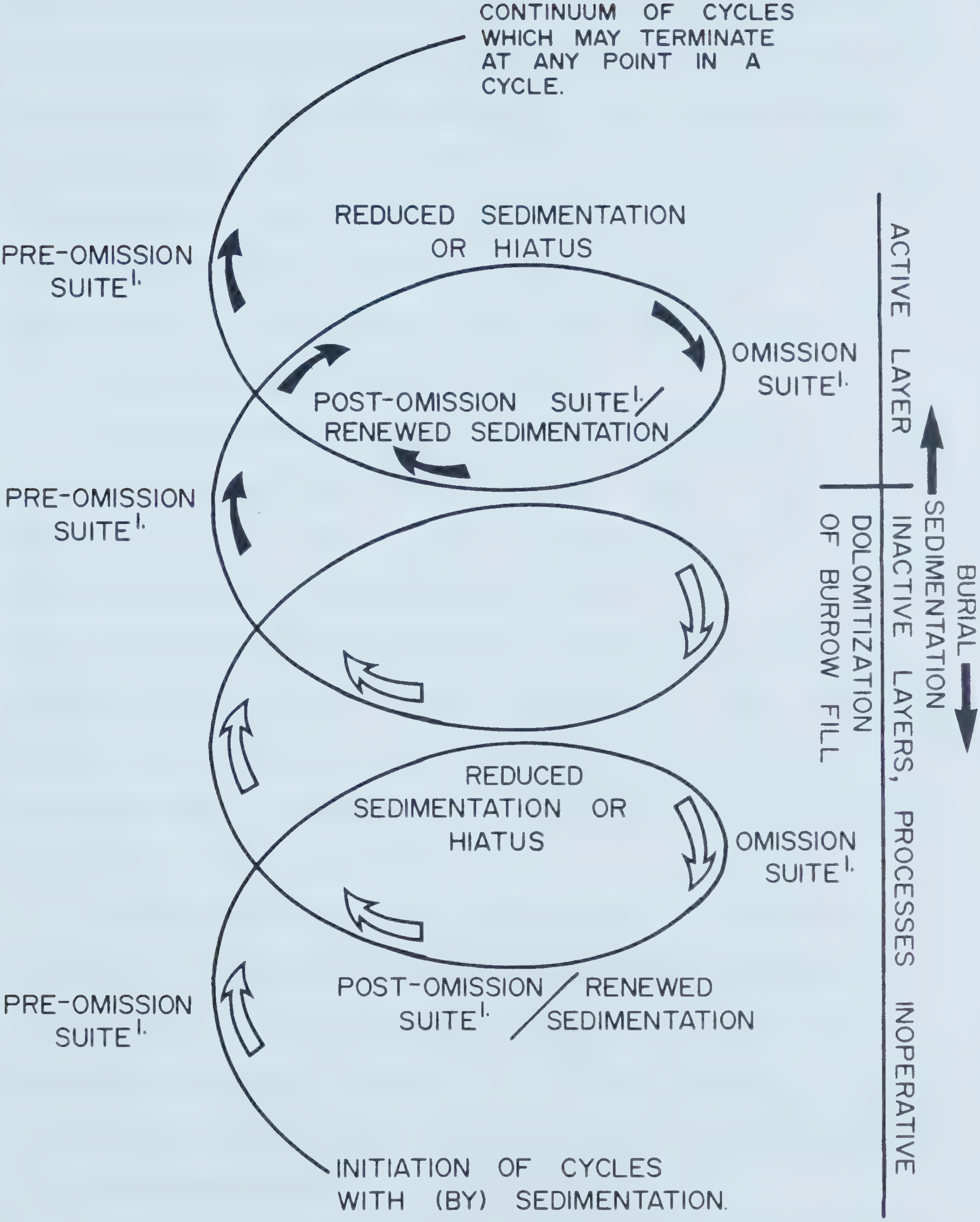
2. Biogenetic controls.
    a) Type of burrowing organism.
    b) Rate of bioturbation.

3. Diagenetic controls.
    a) Rate of early lithification of parent sediment.
    b) Early diagenetic (dolosiltite) fill of burrows.
    c) Later diagenetic processes of burial diagenesis, silicification, pressure solution and recrystallization.

TABLE 5.1. Controls on mottling in dolomitic limestones.

for thick sequences of mottled dolomitic limestones to develop, the processes responsible for their formation were relatively constant through much of the late Ludlovian in the M'Clintock Basin. Deposition of the original sediment was on a relatively stable, shallow carbonate shelf on which sedimentation was slow. Rates of bioturbation were high and distribution of the burrow systems was constrained by early lithification of the sediment. Because of the relationships between the burrows and the early lithification of the sediment, a chronologic succession of burrows relative to the lithification of the sediment is discernible. Subsequent dolomitization of the burrow-fill is considered to be early diagenetic and to have initiated once the other processes had ceased (Fig. 5.3).

The proposed genesis of the mottled dolomitic limestones is similar to that for the nodular chalks of the Upper Jurassic and Cretaceous of England and Denmark as



1. Bromley (1975).

FIG. 5.3. Summary diagram showing the relationship of trace fossil suites (Bromley, 1975) and sedimentation of dolomitic limestones.

described by Fürisch (1973), Kennedy and Juignet (1974), Bromley (1975), Kennedy and Garrison (1975) and Ekdale and Bromley (1984). They recognized that when burrow systems occur in association with omission surfaces or hardgrounds, a chronological succession of burrows is discernible. Bioturbation, in conjunction with early submarine cementation, are mechanisms that were largely responsible for the formation of nodular chalks.

A chronological succession of burrow systems relative to the lithification cf the limestone lumps or layers is present: 1) a prelithification or preomission suite, 2) a synlithification or omission suite and 3) a postlithification or postomission suite (Fig. 5.3). The terminology of Bromley (1975) is applied to this rock suite as the ichnoassemblage and the temporal relationships of the traces to the lithification of the sediment are similar to that of the Cretaceous Chalks.

The consistancy of this ichnocoenose (comprised of *Chondrites*, *Thalassionoides* and *Planolites*) suggests that the environmental conditions remained relatively stable throughout the deposition of the mottled dolomitic limestones. These traces can occur in any of the suites and in varying abundances.

The terminology for disconformities of Kennedy and Garrison (1975) is used in this study: 1) omission surfaces refer to disconformities marked by differences in sediment type across the surface, 2) incipient hardgrounds represent

further cementation of the sediment to form semi-continuous or continuous layers of limestone, 3) erosion surfaces indicate removal of unconsolidated sediment to expose the lithified limestone and 4) hardgrounds which show evidence of exposure as a sea floor. Fürsich (1979) recognized a series of genetic hardgrounds in which all of the above disconformities could be encompassed.

Rates of Sedimentation

The lines of evidence to support slow rates of sedimentation are essentially the same as those supportative of early lithification. Sedimentation was either cyclic such that negative fluctuations in the rate of sedimentation resulted in minor disconformities or slow enough to allow submarine cementation below the sediment-water interface and the development of omission surfaces. Erosion surfaces and hardgrounds were lesser developed and were possibly random events during a period of relatively constant sedimentation.

The presence and nature of the rhodoliths attests to a slow rate of sedimentation. Nonabraded rhodoliths in the micrite lumps suggest slow rates of sedimentation whereas the abraded rhodoliths in the skeletal calcarenites attest to rapid deposition and reworking of the sediment.

Slow rates of sedimentation as a requisite for the formation of nodular or mottled limestones has been postulated by many workers (Bromley, 1967; Shinn, 1969; Kennedy and Juignet, 1974; Jenkyns, 1974; Müller and

Fabricius, 1974; Noble and Howell, 1974; Kennedy and Garrison, 1975; Rubin and Friedman, 1977; Jones *et al.*, 1979; Abed and Schneider, 1980).

A cycle of sedimentation is apparent when the degree of bioturbation is examined relative to the rate of sedimentation. Burrow structures are most common in the omission suite; a period of reduced or interrupted sedimentation whereas burrows of the preomission and postomission suites are less common. Should this represent a true situation rather than one of recognition of burrow structures, the degree of bioturbation and the rate of sedimentation may show an inverse relationship.

Prelithification (Preomission) burrows

The preomission suite represents burrowing activity in the original sediment prior to lithification. *Chondrites* and *Thalassinoides* are present but preservation, and hence recognition, is generally poor due to the similarity between the burrow-fill and the host rock (Plate 23, Fig. 3), erasure by reworking by later burrowers or erosion of burrow systems. Recognition of *Thalassinoides*, in this suite, is predominantly based on relict bioturbation textures such as color mottling in the limestone lumps of layers (Plate 23, Fig. 3).

The best example of preomission burrows are *Chondrites* in the micrite lumps or layers (Plate 22, Fig. 1; Plate 23, Fig. 4; Plate 25, Figs. 1 and 3). These burrows are usually

filled by micrite and are conspicious by their color mottling. *Chondrites* of the preomission suite commonly show truncation by other burrow systems (Plate 22, Fig. 1) and do not transgress disconformities such as omission surfaces.

Some of the larger burrows may have subsequently been reworked by burrowers of later suites or have been truncated during erosion of the unconsolidated sediment overlying the omission surface.

Lithification and omission suite burrows

The early lithification of the sediment and the omission suite of burrows are contemporaneous processes that are largely responsible for the mottled nature of the limestones. The lateral and vertical continuity of the limestone lumps or layers is determined by the relative rates of early lithification versus the rates of bioturbation. The rate of bioturbation is the dominant factor in determining this continuity of the limestone lumps or layers. This is evident in Types I and II which occur as lateral equivalents over distances of less than a metre. It can only be assumed that the rate of lithification and sedimentation would be relatively constant over such a small area, leaving the rate of bioturbation as the unknown variable.

The rates of these processes are only relative; an absolute time framework cannot be established. As such, the rate of bioturbation can imply large numbers of burrowers

over a short period of time or a lesser number of burrowers over a longer time period. This variable, in conjunction with the rates of lithification, could explain much of the nature of the mottled dolomitic limestones. Rubin and Friedman (1977) noted that the degree of bioturbation is controlled by abundance of burrowing organisms and the sedimentation rate. Shinn (1969) also noted the relationship of increased bioturbation in response to decreased sedimentation.

Early lithification

The formation of limestone lumps or layers is attributed to cementation below the the sediment-water interface. This conclusion is based on negative evidence such as the lack of mineralized surfaces, or of borings or an epifauna that are indicative of exposed submarine surfaces. Some of the limestone lumps or layers show evidence of exposure at this interface but these are the exception rather than the norm. There is little evidence to indicate to what degree, if any, the sediment was lithified prior to exposure and there is no evidence to indicate to what depth cementation took place below the sediment-water interface.

The association of early diagenetic lithification of the sediment, at or near the sediment-water interface, and the formation of nodular or mottled limestones has been documented by many workers (Beales, 1953; LeBauer, 1965;

Bromley, 1967, 1975; Hallam, 1967; Shinn, 1969; Rasmussen, 1971; Goldring and Katzmierczak, 1974; Kennedy and Juignet, 1974; Jenkyns, 1974; Noble and Howell, 1974; Kendall, 1977; Jones *et al.*, 1979; Mullins *et al.*, 1979; Abed and Schneider, 1980).

The evidence supporting early lithification (cementation) are: 1) the formation of limestone lumps or layers, 2) the presence of rare encrustations of *Favosites* and stromatoporoids on the limestone lumps, 3) the termination of burrows in the matrix at the margins of the limestone lumps and 4) the presence of hardgrounds or erosion surfaces.

The presence of encrusting organisms on the limestone lumps indicates that the limestone lumps were a fully lithified substrate prior to exposure at the sediment-water interface. Encrustations occur only on the lumps and do not extend laterally into the matrix. This suggests that athe matrix was soft and could not support encrusting organisms whereas the limestone was a hard substrate. The absence of hardgrounds, laterally adjacent to the encrusted lumps, indicates that these surfaces were not maintained as exposed submarine surfaces or as a seafloor. Similar encrustions on nodules in nodular limestones were documented by Jenkyns (1974) and Jones *et al.* (1979).

The termination of burrows at the margins of the limestone lumps and layers reflects the inability of the burrowers to penetrate the lithified limestone (Plate 19,

Fig. 6; Plate 22, Fig. 1). This relationship is most apparent in the dolosiltite matrix where *Chondrites* abut against or follow along the limestone/matrix interface but do not transgress it. A similar relationship between the burrows and the limestone lumps was documented by Jones *et al.* (1979) in the rubbly mottled dolomitic limestones of the Read Bay Formation on Somerset Island.

True hardgrounds and/or erosion surfaces are rare and are more commonly associated with Type III mottled dolomitic limestones whereas omission surfaces are more commonly associated with Types I and II mottled dolomitic limestones. The distribution of these surfaces in the mottled dolomitic limestones further substantiates the proposal that Type III mottled dolomitic limestones formed under slightly different sedimentological conditions than did Types I and II. Erosion surfaces dominate over hardgrounds and are evident by erosion and truncation of fossils (Plate 10, Fig. 2; Plate 24, Figs. 1,2 and 3). Hardgrounds were recognized by the presence of mineralized surfaces; borers and an epifauna are rare to absent.

The Jurassic and Cretaceous chalks differ from the strata of this study in that the latter lack well-developed hardgrounds. This lack of hardgrounds, however, may be a function of observation rather than absence. Sarkar *et al.* (1980) noted the problems of hardground recognition in limestone sequences other than soft chalks. Several factors may account for the difficulty in recognition of

hardgrounds.

One is the rubbly nature of Types I and II mottled dolomitic limestones. The rubbly nature may mask the presence of hardgrounds and may be one reason why the incidence of hardgrounds and erosion surfaces is higher in Type III mottled dolomitic limestones than the other types.

A second factor is a time factor. The duration of submarine exposure may not have been long enough for the indicative criteria to develop. Fürsich (1978, 1979) recognized that a time factor was necessary for the development of a climax community and/or of a mineralized surface in association with hardgrounds. While mineralized hardground are present in the Upper Silurian strata (Plate 23, Figs. 1 and 2; Plate 24, Fig. 4), the presence of hardgrounds where erosion was important must be considered. These erosion surfaces are planar (Plate 24, Figs. 1, 2 and 3) and represent postlithification erosion of a hardground. Similar eroded hardgrounds in the Jurassic chalks of southern England were documented by Fürsich (1979).

A third factor is the evolution of boring organisms. Kobluk *et al*. (1978) documented the increase in diversity of macroborers during the Paleozoic. The Silurian contained relatively few macroborers of which only *Trypanites* was recognized. The poorly defined hardgrounds in the Upper Silurian can possibly be attributed to a combination of the above factors.

Synlithification or omission burrows

The omission suite represents burrowing activity that occurred contemporaneously with the lithification of the sediment. In some cases, lithification may have been initiated prior to the bioturbation. *Thalassinoides* dominates the omission suite (Fig. 5.1). The increase in the density of burrow systems associated with the development of omission surfaces and early lithification may reflect a decrease in the sedimentation rate; hence, the cyclicity of sedimentation may be real rather than apparent. The dependence of the degree of bioturbation on sedimentation rates was documented by Shinn (1969) and Rubin and Friedman (1977). The resulting pattern of burrows is partly controlled by the "imposed horizontality" (Bromley 1967, 1975) on the burrowing organisms by the continuity of the limestone lumps or layers.

In the case of Type I mottled dolomitic limestones, imposed horizontality by the limestone lumps was minimal. Type II represents the situation in which the lateral expansion of the limestone lumps to form incipient hardgrounds (Kennedy and Garrison, 1975) placed some restriction of the vertical movement of the burrowing organisms. The greatest restriction on burrowing organisms is in Type III mottled dolomitic limestones in which the incipient hardgrounds or simple hardgrounds restricted burrowers to horizontal systems except where breaches in the limestone layers previously existed.

Burrows of the omission suite transgress the discontinuities, suggesting bioturbation initiated prior to the lithification or the reworking of previous burrow systems. The omission suite is obvious by the density of burrows (Plate 25, Fig. 4; Fig. 5.2), their crosscutting relationships and the early diagenetic dolomite fill (Plate 19, Fig. 4).

Postlithification (postomission) burrows

Chondrites, *Thalassinoides*, *Planolites* and *Palaeophycus* occur in this suite. *Chondrites* occur in *Thalassinoides* as either distinct burrow systems or as color mottling in the dolosiltite matrix (Plate 19, Fig. 6; Plate 22, Fig. 1; Plate 26, Figs. 2 and 3). The limit of these burrows is circumscribed by existing burrows and the limestone lumps and layers. *Chondrites* in other burrows has been documented by other workers (Bromley, 1967, 1975; Shinn, 1968b; Kennedy, 1970; Kendall, 1977; Bromley and Ekdale, 1984; Ekdale and Bromley, 1984).

The postomission suite marks the renewal of sedimentation (Bromley, 1975; Kennedy and Garrison, 1975) or the abandonment of existing burrows and the upward migration of the burrowers as a result of lithification below the omission surface and the presence of unconsolidated sediment above the omission surface. As such, the postomission suite of one cycle may, in part, be the preomission suite of the successive cycle. Delineation of the two suites may be

impossible. This overlap of cycles accentuates the fact that the processes responsible for the formation of mottled dolomitic limestones are continuous processes and not discrete entities. Burrowers of the postomission suite may modify pre-existing burrows by reworking of the burrows.

Other burrows of the postomission suite are commonly filled by micrite suggesting the fill originated from the new sediment forming above. This may be either the passive fill of empty burrows or the reworking of existing burrows. The presence of a micrite fill may also represent the timing of the fill. If the early diagenetic dolomite is considered to be essentially contemporaneous with the cessation of early lithification of the sediment and of bioturbation; this burrow-fill represents either a burrow-fill after early dolomitization or a permeability barrier such that the original fill was preserved.

Early diagenetic dolomite

Dolomitization of the burrow-fill may be essentially a contemporaneous process, initiated upon cessation of a cycle of sedimentation, bioturbation and lithification but occurring during successive cycles (Fig. 5.3). The dolomite is restricted to the burrows with virtually no dolomitization of the bioclastic material in the burrows or of the limestone lumps or layers. Morrow (1978) and Jones *et al*. (1979) attributed the pattern of dolomitization to the increased porosity and permeablilty of the matrix versus the

lithified lumps. This increased porosity and permeability is also evident by the preferential movement of copper-rich solutions through the dolomitized burrows. Morrow (1978) and Abed and Schneider (1980), as does this study, attributed the pattern of dolomitization in mottled dolomitic limestones to burrow systems.

Dolomitization had minimal effect on textures existing in the burrow-fill; the bioclasts remain undolomitized and the *Chondrites* are preserved although dolomitized. There is no petrographic evidence to suggest dolomitization occurred prior to the postomission suite; rather petrography suggests that dolomitization occurred postbioturbation. The uniformity of the dolomite throughout the sequence of mottled dolomitic limestones suggests that the mode of dolomitization remained constant.

The origin of the dolomitization fluids is unknown and any discussion of their origin would be speculative. Perhaps, extensive geochemical analysis may offer some answers as to the origin of the dolomitization but that is outside the realm of this study. It is possible that the thick sequence of lower Paleozoic dolostones, underlying the mottled dolomitic limestones, may have supplied magnesium ions for the dolomitization fluids. Kendall (1977) and Jones *et al.* (1979) speculated on the origin of the dolomitization in mottled dolomitic limestones but did not reach a conclusion. Morrow (1978) attributed initial dolomitization of the burrow-fill sediments to salinity fluctuations in a

shallow lagoon. He considered that this contemporaneous dolomite served as a nuclei for further dolomite precipitation during later diagenesis. This may be a possible mechanism as evidence of seasonal salinity fluctuations occurs in the dolostones of the Cape Storm and Somerset Island formations.

E. Diagenesis

Two phases of diagenesis have already been discussed; early lithification of the parent sediment and early diagenetic dolomitization of the burrow-fill. Late diagenesis was not a major factor in the formation of the mottled dolomitic limestones and includes only processes that altered existing mottled dolomitic limestones. Locally, chert (flint) replaced the dolosiltite matrix of the burrow-fills (Plate 20, Fig. 2; Plate 23, Fig. 3). Bromley (1967), Kennedy (1970), Rasmussen (1971), Kennedy and Juignet (1974) and Kennedy and Garrison (1975) considered the flint in the Cretaceous Chalks to be early diagenetic in origin.

The cherts are light brown-grey to buff on fresh and weathered surfaces. Silica may replace dolomite in some of the rhombs or totally replace the matrix, indicating that the cherts formed after the dolomitization of the burrows. Fibrous chalcedony occurs primarily as a replacement mineral of calcite in the bioclasts and is probably associated with weathering at the present day surface. Silicified fossils

appeared to be most common on weathered exposures and rare on fresh outcrop. In units Q-30 and 31, the brachiopods contained quartz crystals, forming small geodes.

The is little or no evidence of compaction in either the limestone lumps and layers or in the dolosiltite matrix. The only evidence of compaction is the presence of rare fractured and crushed bioclasts in both lithologic components. Sarkar *et al.* (1980) considered deformed allochems as evidence of hardgrounds or emersion surfaces. The timing of the bioclast deformation is unknown and given the problem of hardground identification in the mottled dolomitic limestones, this may be another criterion for recognizing hardgrounds in this strata.

Late diagenetic dissolution occurred as a result of lithostatic pressures and tectonic pressures. In the dolsiltite matrix, microstylolite swarms (Wanless, 1979) or flaser structures (Kennedy and Garrison, 1975; Garrison and Kennedy, 1977) occur and may be laterally continuous with stylolites in the micrites (Plate 23, Fig. 3). These solution seams are generally horizontal in response to lithostatic pressures or oblique in response to tectonic pressures.

The mottled dolomitic limestones may be dolomitized or locally dolomitized. This is more common in the Somerset Island Formation than either the Cape Storm or Douro formations. In some cases, the limestone is dolomitized with little or no effect on the matrix (Plate 21, Fig. 1) and

ghost textures in the lumps may be recognized (Plate 28, Figs. 1, 2, 3 and 4).

Dedolomitization

Dedolomitization is a late diagenetic process in the mottled dolomitic limestones. The distribution of the dedolomites in the Upper Silurian strata on eastern Prince of Wales Island is largely controlled by the distribution of the mottled dolomitic limestones. Dedolomitization refers to the process in which dolomite is replaced by calcite. The term *calcitization* (Swett, 1965; Smit and Swett, 1969) has been proposed as a more acceptable term for calcite-after-dolomite but the term *dedolomitization* is recognized as having priority in the literature and is thus retained.

The dedolomites occur predominantly in the dolosiltite matrix of the burrow-fills but can also occur as isolated rhombs in the micrite lumps. The predominance of dedolomites in the matrix is attributed to: 1) a higher proportion of dolomite in the matrix and 2) the increased permeability and porosity of the matrix which allows for greater movement of the dedolomitization fluids. Morrow (1978) noted that the burrows in the Upper Ordovician mottled dolomitic limestones on Devon Island were selectively dolomitized due to greater diffusion of magnesium from seawater in the more porous burrowed sediments. The corollary for dedolomitization in the Upper Silurian mottled dolomitic limestones is proposed.

The association of increased dedolomitization with increased porosity has also been documented by Al-Hashimi and Hemingway (1973). Dedolomitization of dolomite rhombs in the dolomite matrix of the rubbly dolomitic limestones of the Read Bay Formation on Somerset Island was also documented by Jones *et al*. (1979).

The dedolomites in the Upper Silurian mottled dolomitic limestones on eastern Prince of Wales Island are attributed to surface weathering. This is probably related to Recent weathering but may also represent relic weathering as the Upper Silurian strata were exposed in the earliest Devonian in response to Pulse 2 of the Cornwallis Disturbance (Kerr, 1977). Most workers have attributed dedolomitization to near-surface weathering, whether relic or Recent (Schmidt, 1965; de Groot, 1967; Evamy, 1967; Goldberg, 1967; Katz, 1968, 1971; Braun and Friedman, 1970; Chafetz, 1972; Al-Hashimi and Hemingway, 1973; Frank, 1981; Kuslanksy and Friedman, 1981). Budai *et al*. (1984), however, recognized dedolomites in the Western Overthrust Belt of Wyoming and considered the dedolomites to reflect high temperature burial conditions resulting from tectonic thrusting. The dedolomites in the Upper Silurian mottled dolomitic limestones may, in part, be burial dedolomites.

Initially, the presence of dedolomites in the strata of this study were thought to be associated with a disconformity, coincident with the Douro Formation-Somerset Island Formation boundary. Several criteria supported this

contention: 1) the lateral and vertical distribution of the dedolomites, 2) the stratigraphic range of red coloration in the mottled dolomitic limestones, 3) the local variability in the thickness of the Somerset Island Formation, 4) the loss of section in the Douro Formation and 5) the abrupt changes in the lithological and paleontological succession across this boundary. The latter three criteria are attributed to structural controls whereas the former two are associated with surface weathering.

Dedolomites are most abundant (and evident) in the southern portion of the study area (Fig. 5.4) but occur in the mottled dolomitic limestones throughout the study area. The stratigraphic range of the dedolomites (Fig. 5.5) is best demonstrated on central-eastern Prince of Wales Island because of the greater abundance of dedolomites and the density of measured sections. The upper limit of the dedolomites is approximately coincident with the Douro Formation-Somerset Island Formation boundary (Fig. 5.5) and reflects the change in lithology from the mottled dolomitic limestones to planar-bedded dolostones and limestones across this boundary. The contrast in lithology and paleontology across this boundary is associated with a shallowing of the M'Clintock basin in response to a minor tectonic pulse of the Boothia Horst in the late Ludlovian.

A pronounced red coloration occurs in some of the bioclasts in the mottled dolomitic limestones, particularly when the bioclast is infilled by sparry calcite. This

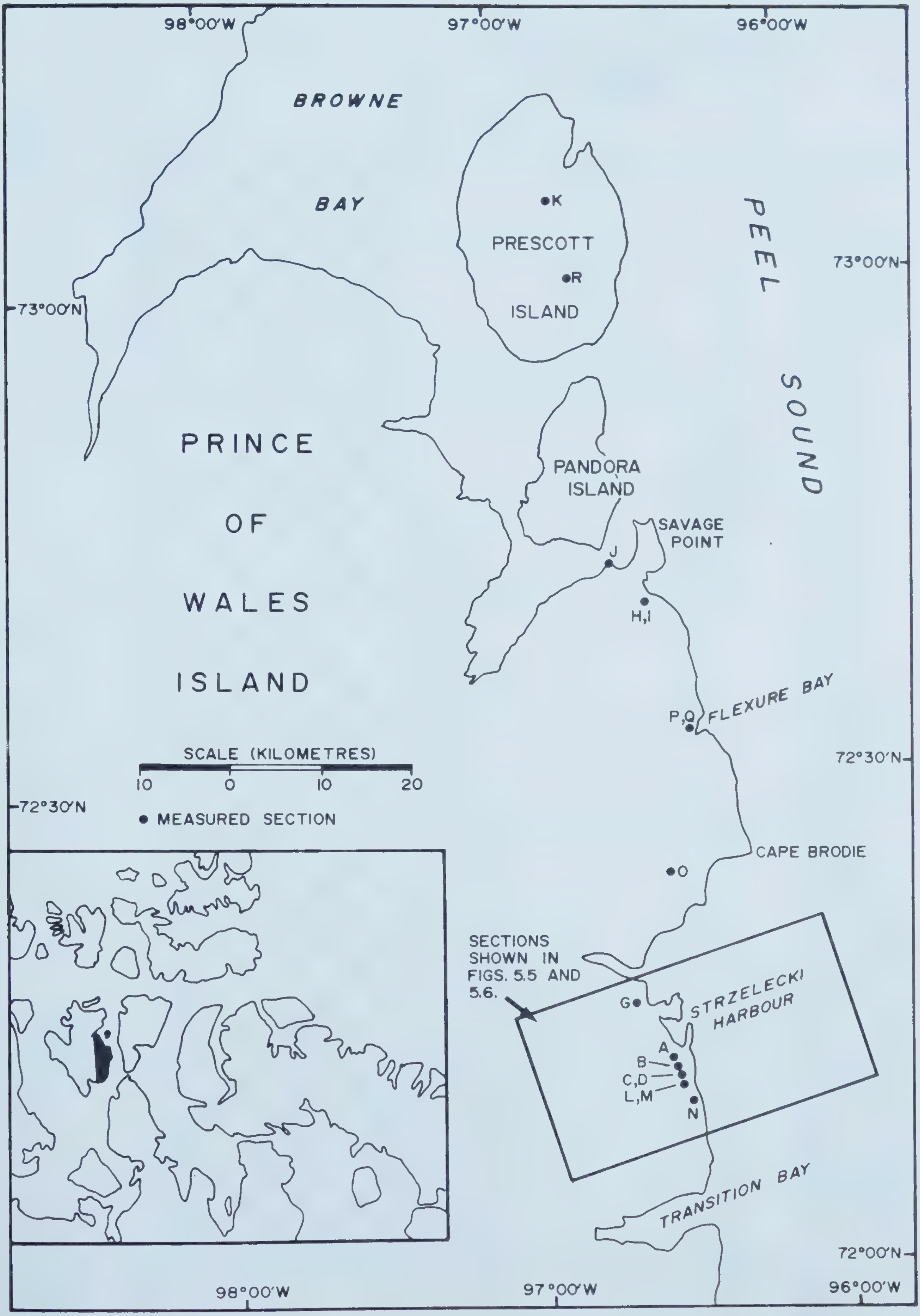


FIG. 5.4. Location of measured sections on Prince of Wales and Prescott Islands.

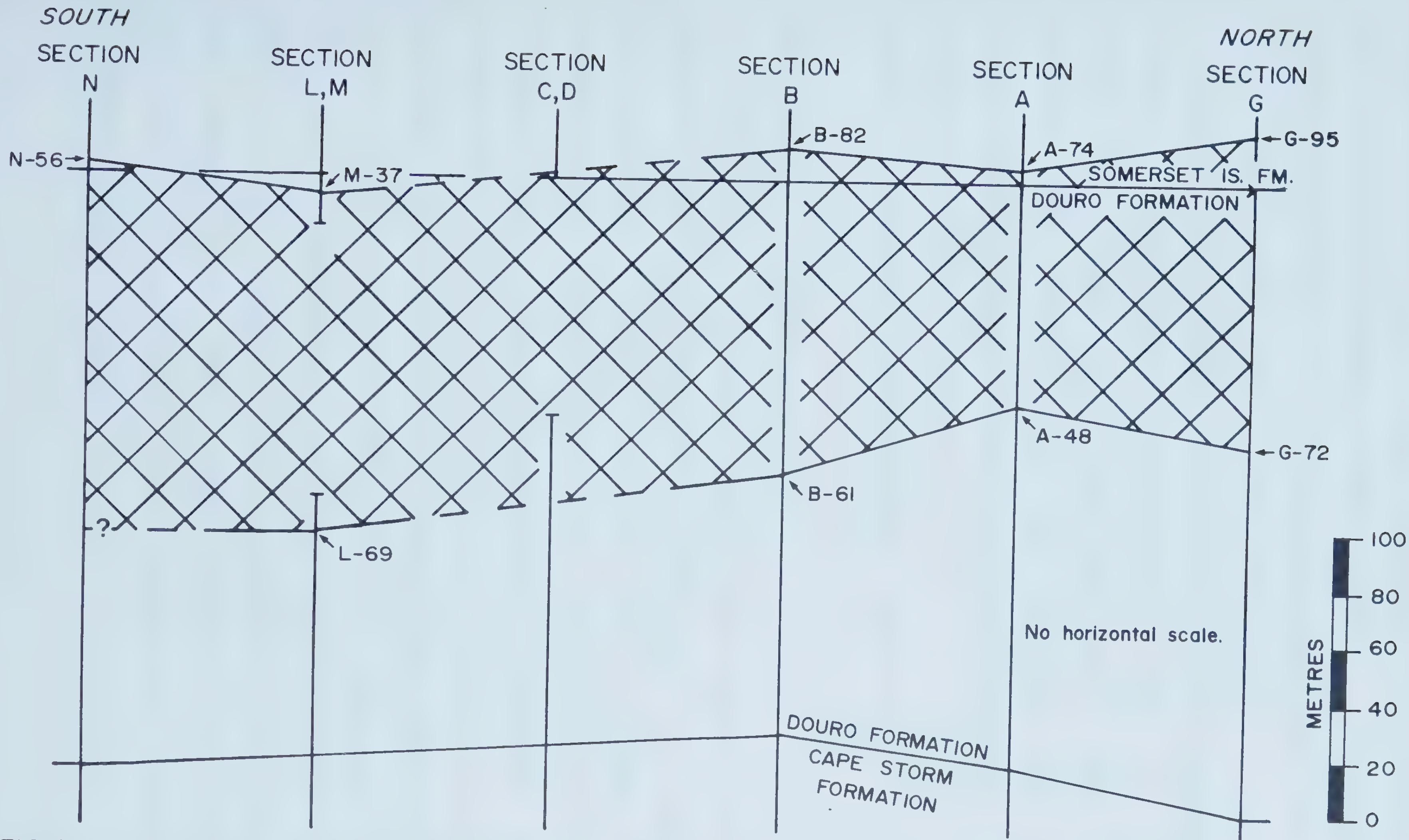


FIG. 5.5. Stratigraphic distribution of dedolomites in the Upper Silurian strata on central-eastern Prince of Wales Island. Datum is the base of the Somerset Island Formation. Sections C and L are set on the base of the Douro Formation to approximate the thickness of missing section.

coloration shows a similar stratigraphic distribution (Fig. 5.6) as the dedolomites and the coloration is presumed to indicate the presence of ferruginous compounds. Cotter (1966) recognized a similar red-orange calcite in the molds of organisms and cavities of the Mississippian Lodgepole Formation in Montana. The red coloration indicated the presence of montmorillonite clays and Cotter (1966) attributed the coloration to a relic weathered zone. As with the dedolomites, the timing of the red coloration in the mottled dolomitic limestones is of question and may be related to both relic and/or Recent weathering.

Petrology of the dedolomites

Dedolomites can assume several petrographic textures; a list of criteria for recognizing dedolomites is presented in Shearman *et al*. (1961), Evamy (1963, 1967), Goldberg, (1967) and Al-Hashimi and Hemingway (1973). The staining technique of Evamy (1963), using a combined stain of alizarin red-S and potassium ferricyanide in a single solution, was applied to all the thin sections. As a check on the staining method, 20 thin sections were stained using only a potassium ferricyanide solution. Both staining methods showed ferroan calcite and/or ferroan dolomite to be absent from these rocks. Use of a Kevex 7000 energy dispersive x-ray analyzer also showed ferroan dolomites or any other iron compound to be absent from these samples.

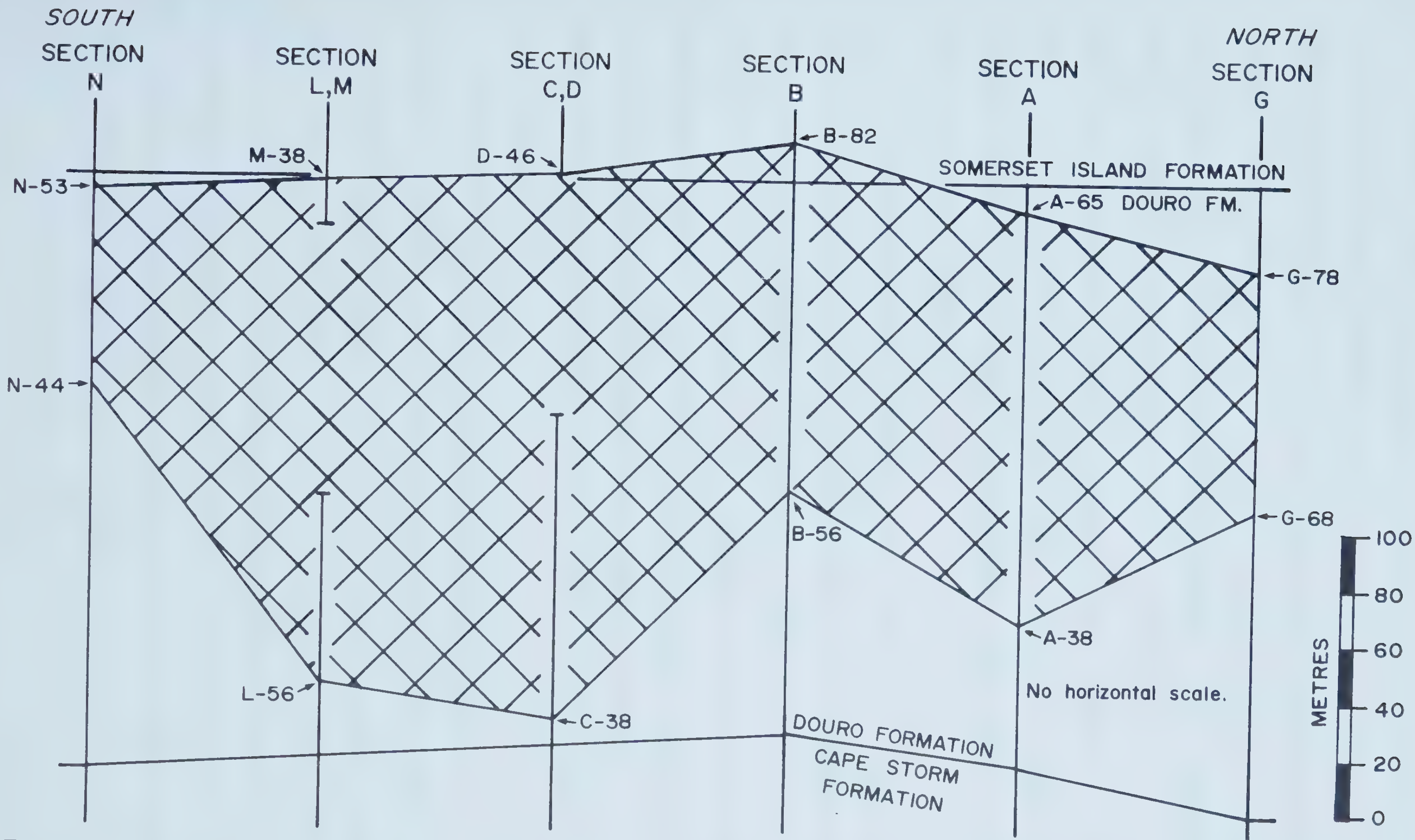


FIG. 5.6. Stratigraphic distribution of red-colored bioclast infill in the Upper Silurian strata on central-eastern Prince of Wales Island. Datum is the base of the Somerset Island Formation. Sections C and L are set on the base of the Douro Formation to approximate the thickness of missing section.

Several dedolomitization textures are present in the mottled dolomitic limestones: 1) relict rhombs which are partially replaced by calcite. They may occur as zoned rhombs (Plate 29, Figs. 3 and 4), rhombs with a calcite core or rhombs with corrosion along a crystal face (Plate 30, Figs. 5 and 6) and 2) rhombs totally replaced by polycrystalline calcite (Plate 29, Figs. 1 and 5; Plate 30, Figs. 1 and 3). The replacement calcite may be either coarse or fine grained and either texture may show centripetal or centrifugal replacement.

In most examples, the dolomite rhombs show only partial replacement; having either a calcitic core or embayments. The calcite may mimic the limestone texture (Plate 29, Figs. 1 and 2; Plate 30, Figs. 1, 2, 3 and 4). Where the rhombs have been totally replaced by calcite, only a ghost of the original rhomb remains (Plate 29, Figs. 1 and 5). Cross-polarized light accentuated such rhombs (Plate 29, Figs. 2 and 6).

The dedolomites of this study do not show an association with iron oxides or hydroxides as documented by other workers (Shearman *et al.*, 1961; Evamy, 1963; Katz, 1968, 1971; Wolfe, 1970; Al-Hashimi and Hemingway, 1973; Frank, 1981). These are generally considered to be bi-products from the dolomitization of ferroan dolomites. The absence of ferroan dolomites may explain the absence of iron oxides and hydroxides.

Mechanisms of dedolomitization

The mechanism postulated for dedolomitization is the reaction of dolomite with a calcium sulphate solution derived from either the dissolution of gypsum or the oxidation of pyrite (Lucia, 1961; Evamy, 1963; de Groot, 1967; Goldberg, 1967; Folkman, 1969; Chafetz, 1972; Lucia, 1972; Al-Hashimi and Hemingway,1973; Warrak, 1974; Back *et al.*, 1984). Kastner (1982) and Back *et al.* (1984) recognized that it is the calcium in the calcium sulphate solutions that controls replacement.

The absence of gypsum or of gypsum pseudomorphs in the lower Paleozoic strata on eastern Prince of Wales Island rules out this mechanism for dedolomitization. Pyrite and hematite occur as accessory minerals in the dolosiltite matrix which may support oxidation of pyrite as a mechanism for dedolomitization. However, if oxidation of pyrite was a mechanism for dedolomitization; it would be expected that the patterned dolostones which contain oxidized pyrite, would show evidence of dedolomitization. The absence of dedolomitization textures in the xenotopic patterned dolostones may be a function of the low permeability and porosity of these rocks.

Whatever the mechanism, the dedolomites are associated with the mottled dolomitic limestones, particularly the dolomitized burrow-fills. It is a post-depositional diagenetic process and is most probably Recent rather than relic. The distribution of the dedolomites would appear to

be a function of the porosity and permeability of the rocks and the nature of the interstitial fluids.

F. Summary

The synthesis proposed for the genesis of the mottled dolomitic limestone of this study appears to satisfy the characteristics of these rocks: 1) the discrete nature of the two lithologic components, 2) the distribution of the limestone lumps and layers versus the dolosiltite matrix and 3) the thick succession of these rock types. While depositional and diagenetic controls are important, the rate and nature of the bioturbation appears to be a dominant controlling factor on the variations within the spectrum of mottled dolomitic limestone.

When examined in chronological succession, it is the conjunction of early lithification and the omission suite of burrowers that are largely responsible for the nature of the rocks as evident in the present. The pattern of the burrow systems and/or of the limestone component relies on the restriction imposed by one process on the other and visa versa. This is most apparent in Type III mottled dolomitic limestone in which the lateral continuity of the limestone layers restricted the burrowers to horizontal systems. Types I and II show a lesser degree of imposed horizontality (Bromley, 1967, 1975) of the burrow systems. Early dolomitization appears to have been selective; being almost exclusively confined to burrow systems of the omission site.

The poor quality of preservation of the burrow systems, predominantly the larger burrows such as *Thalassinoides*, *Palaeophycus* and *Planolites* has been a hinderance to the development of this synthesis. Fortunately, there was excellent preservation in some units throughout the succession; without which recognition of burrow systems of lesser quality would have been impossible. It is almost certain that without the detailed work in the Cretaceous Chalks of Bromley (1967, 1968, 1975), Kennedy (1967) and Kennedy & Garrison (1975), this synthesis would be lacking. The similarities between the biogenetic and sedimentological structures occurring in the strata of this study and the chalks suggests that the genesis of the two, barring changes in the geologic setting, are akin. Perhaps the largest opponent of applying their model is the lesser degree of preservation of biogenetic structures in the strata of this study.

The mottled dolomitic limestone of the Upper Silurian strata on eastern Prince of Wales Island share a link through time and genesis, extending from the Cape Storm Formation to the Somerset Island Formation. The time interval represented by these rocks indicates that a stable carbonate platform with little change in paleoenvironmental conditions dominated. These rocks show a consistancy in lithologic and biogenetic processes in both a spatial and temporal context.

VI. REGIONAL FACIES OF THE UPPER SILURIAN

A. Introduction

Aspects of both the lithofacies and the biofacies are examined in this chapter. The emphasis on lithofacies reflects the lack of well-defined biofacies in the strata of this study. In the mottled dolomitic limestones, a well-developed zonation of the brachiopod genus *Atrypoidea* occurs. Other biofacies such as the stromatolites and the ichnotaxa are organosedimentary structures and because of their role in the formation of specific rock types are discussed with the lithofacies.

The lithofacies are discussed in an interpretative sense and used to establish the paleogeography of the study area. While some reference to modern analogues is employed, the interpretation of the lithofacies is largely based on criteria evident in the strata (Fig. 6.1) rather than by modern analogue. Playford and Cockbain (1976, p. 409) recognized the problem with the application of modern analogues and approach the application with caution:

> "There are lessons to be learned from the Hamelin Pool stromatolite story regarding the use of modern models to interpret ancient rocks. It shows the danger of accepting incompletely known modern environments as definitive guides to the past. The nature of ancient environments should be deduced from evidence in the ancient rocks themselves, without the mandatory need for modern analogues."

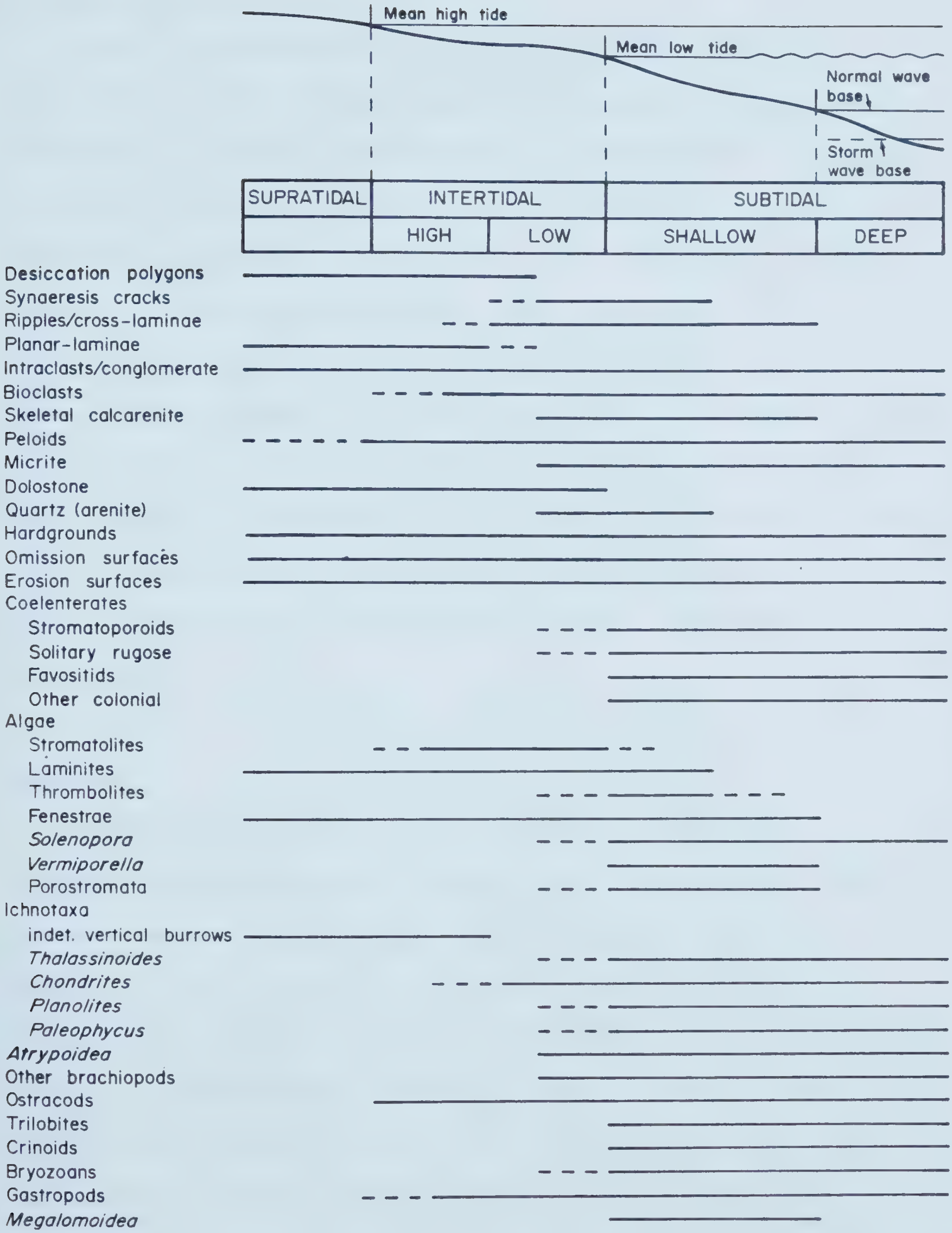


FIG. 6.1. Paleobathymetric range of sedimentological and paleontological criteria occurring in the Upper Silurian on eastern Prince of Wales Island.

B. Cape Storm Formation

The Cape Storm Formation includes three informal members: 1) member A comprised of dolostone and quartzose dolostone, 2) member B comprised of stromatolitic dolostone and algal laminated dolostone and 3) member C comprised of dolostone and dolomitic limestone. The members are highly variable in thickness throughout the study area (Fig. 6.2). Some of the lithofacies (ie. quartzose dolostone) and the members (ie. stromatolitic dolostone) are not present in all the sections but generally the members show superposition. The variation in thickness of the Cape Storm Formation (Fig. 6.3) is considered to be due to original depositional differences and not to post-depositional structural modification.

Member A

Description: This member is characterized by dolosiltite and quartzose dolosiltite; the only difference between the two lithologies being the quartz content. Typically, the dolosiltites show characteristics of particulate material such as graded bedding (Plate 31, Figs. 3 and 4), ripple marks (Plate 32, Fig. 3) and low angle cross-laminae (Plate 31, Fig. 1). Other sedimentological features are planar-laminae, desiccation polygons and intraclasts (Plate 5, Fig. 1). The strata contain rare bioclasts and indeterminate vertical burrows (Plate 27, Fig. 3; Plate 31, Figs. 2 and 4).

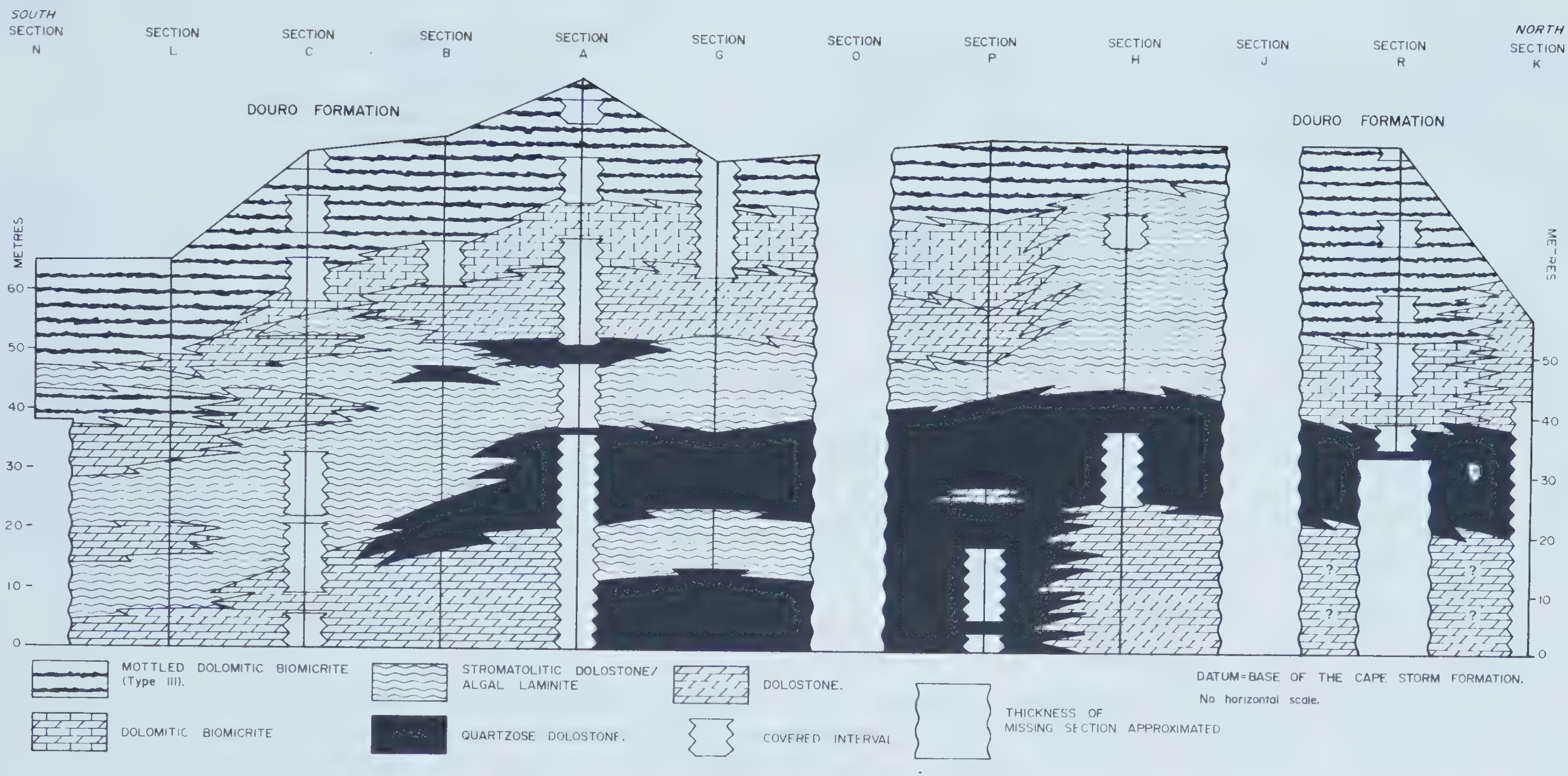


FIG. 6.2. Generalized regional lithofacies of the Cape Storm Formation on eastern Prince of Wales Island.

	CAPE STORM FORMATION Thickness (metres)	DOURO FORMATION Thickness (metres)	SOMERSET ISLAND FORMATION Thickness (metres)
SECTION A	96.6	197.3	178.0
SECTION B	86.5	194.1	167.1
SECTION C,D	83.7	115.4 F	156.8
SECTION G	82.3	241.8	160.6 (+)
SECTION H,I	85.6	174.2	100.3 (+)
SECTION J	—	194.8	84.8 (+)
SECTION K	56.5	101.5	167.3
SECTION L,M	65.4	102.9 (87.3 + 15.6) F	134.3
SECTION N	26.6 (+)	208.4	257.4
SECTION O	—	145.8 (+)	—
SECTION P,Q	86.2	170.3 (63.4 + 106.9) F	195.7
SECTION R	52.6 (+)	262.0	86.5 (+)

FIG. 6.3. Formation thickness in measured sections. (+) indicates measured interval is only part of the formation and one formation boundary is not evident. F indicates faulting within a formation is evident and both formation boundaries were observed.

The dolosiltites are predominantly planar-laminated; the incidence of ripple marks and cross-laminae increases up-member. Graded bedding is common in the planar-laminae, grading from coarse dolosilts and dolosands at the base to fine or medium dolosilts at the top. Typically, the laminae are less than 3 mm thick. Graded dolosilts with a similar lamination thickness were documented by Shinn (1983) in the supratidal beach ridges on Andros Island in the Bahamas.

Blatt *et al.* (1980) and Shinn (1983) considered that the preservation of laminae is inversely related to the activity of burrowing organisms and that there is a relationship between the rate of sedimentation and the rate at which organisms disrupt the sediment. Blatt *et al.* (1980) recognized four environments conducive to the preservation of laminae in carbonate sediments: 1) environments that lack water circulation and therefore lacking oxygen, 2) playa lakes, 3) supratidal environments and 4) algal binding of sediments. Only the latter two environments are recognized in the strata of this study. Algal bound strata do not occur in member A.

The presence of sedimentary structures in this member suggests that the dolosiltites may be largely detrital in origin. Detrital dolomites have been documented by many workers (Sander, 1936; Rodgers, 1954; Deffeyes and Martin, 1961; Amsbury, 1962; Sabins, 1962; Lemon and Blackadar, 1963; Bluck, 1965; Lindholm, 1969; Shinn, 1973, 1983; Dixon and Jones, 1978; Gibling, 1978; Freeman *et al.*, 1983).

In this study, dolomites which show evidence of mechanical deposition will be considered detrital, regardless of whether the dolomites originated from eroded source rocks outside the depositional basin or as locally derived dolomites grains formed within the depositional basin. Sabins (1962) differentiated between detrital dolomites derived from eroded source rocks and primary dolomites which are locally derived clastic grains. This distinction between local and terrigenous provenance was not evident from the petrographic methods used in this study and hence, is not employed.

Perhaps a criteria that may prove useful in distinguishing provenance may be the presence or absence of quartz silts in the dolostones. Locally derived intrabasinal detrital dolomites might be more homogenous than the transported dolomites which may contain quartz silts. Based on this premise, some of the dolostones of member A may be locally derived detrital grains rather terrigenous. Freeman *et al.* (1983) considered detrital dolomites to be terrigenous by their association with terrigenous quartz. They found a correlation between the composition of the Miocene detrital dolomites of Menorca, Spain and the percentage outcrop of Jurassic and Triassic dolostones (Freeman *et al.*, 1983).

Detrital dolomites occur in the Cape Storm and Somerset Island formations. Dolosiltites dominate in both formations. Dolarenites generally occur as lenses or thin beds and are

more common in the Somerset Island Formation than the Cape Storm Formation. Dololutite and dolorudite are rare in both formations.

The dolomites grains are commonly associated with clastic grains, predominantly quartz but rare muscovite and feldspar grains are present. Typically, the quartz grains are coarser than the dolomite grains (Figs. 6.4 and 6.5). This difference in grain size is primarly attributed to the differences in density between the dolomite (2.86) and quartz (2.65) grains.

The dolomite grains show characteristics of detrital clastic grains: 1) syntaxial overgrowths (Plate 33, Figs. 1, 2 and 5; Plate 34, Figs. 2, 3 and 4), 2) polycrystalline grains (Plate 34, Fig. 1) and 3) rimming of opaques along the origin grain margin (Plate 33, Fig. 1).

Syntaxial overgrowths are recognized by the characteristics of the original grain. Typically, the grains are clouded with impurities (Plate 33, Figs. 1, 2, 3 and 4; Plate 34, Figs. 3 and 4) or have a rim of opaques along the grain margin (Plate 33, Fig. 1), whereas the overgrowths are clear. The grains are rounded to slightly rhombohedral in shape, commonly appearing as rhombohedrons with "chipped" corners (Plate 34, Figs. 3 and 4). The overgrowths form rhombohedrons (Plate 34, Figs. 3 and 4) and may develop idiotopic to xenotopic textures.

Discussion: Member A is predominantly supratidal to high intertidal in origin. The lack of evaporitic minerals

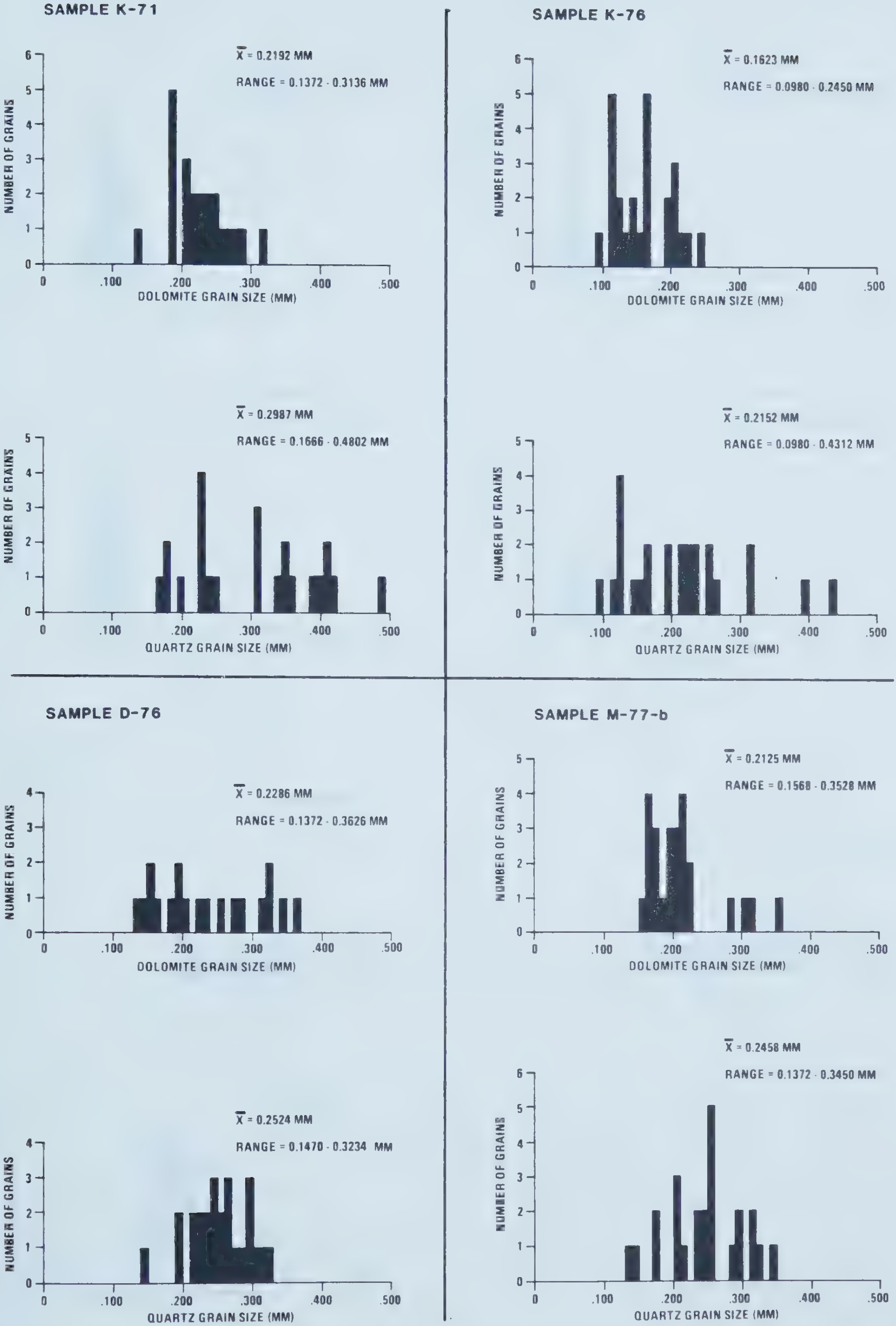


FIG. 6.4. Maximum apparent dolomite and quartz grain size in quartzose dolostones.

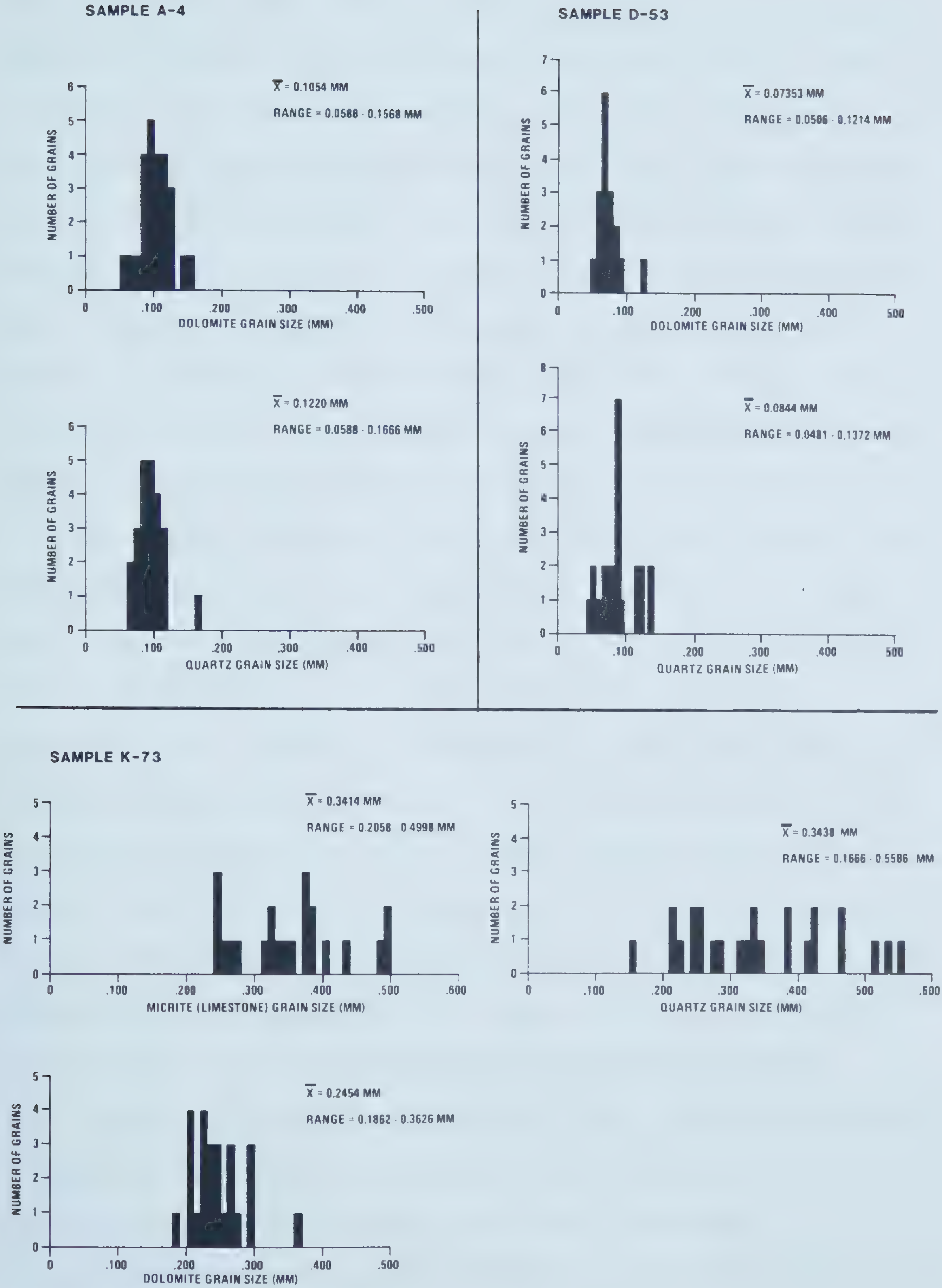


FIG. 6.5. Maximum apparent dolomite, micrite and quartz grain size in quartzose dolostones.

and of abundant desiccation features indicates the strata possibly formed on a humid tidal flat rather than an arid tidal flat. The lack of evaporitic minerals, however, does not suggest that hypersaline conditions were never present but may indicate seasonal variations in the moisture levels and the dissolution of the evaporitic minerals during the "wet" season. A seasonal fluctuation in salinity and the removal of evaporite minerals was reported by Shinn (1983, p. 199) on Andros Island in the Bahamas and by Muir *et al.* (1980) in the Coorong Basin in Southern Australia.

The planar-laminated detrital dolomites of member A may have been deposited in the supratidal to high intertidal zone. Laporte (1967) and Lucia (1972) considered thinly laminated carbonates to be indicative of supratidal deposition. The absence of mudcracks in the supratidal, planar-laminated sediments on Andros Island was documented by Hardie and Ginsburg (1977). They considered periodic wetting was necessary for desiccation features to develop. Shinn (1983), however, documented detrital dolomites in the Persian Gulf and suggested that they indicate a more arid environment. If a seasonal fluctuation in the moisture levels exists, member A may represent the deposits preserved only during the humid season and the evaporitic minerals formed during the arid season may have undergone penecontemporaneous dissolution. This would also support the lack of other criteria indicating arid conditions.

Alternatively, it may be suggested that the dolostones composed of transported dolomite grains may not show the same degree of desiccation as *in situ* developed dolostones. This may be due to the reworking of the sediments by deflation. Shinn (1973) noted that desiccation features on supratidal flats in the Persian Gulf may be rare to absent due to deflation; making the distinction between the supratidal and intertidal deposits impossible.

The source of much of the dolomite is probably reworked eolian or fluvial transported sediments. Exposed lower Paleozoic and Proterozoic strata along the Boothia Horst could have supplied both dolomite and quartz grains. During deposition of members A and B of the Cape Storm Formation, the Boothia Horst was possibly a low-lying emergent feature.

Member B

Description: Member B is composed of variably quartzose dolosiltites and the presence of recognizable algal structures distinguishes it from member A. The two lithofacies, stromatolitic dolostone and algal laminated dolostone, occur as a complex of interdigitated lenses and beds. Member B could be considered a biofacies because of the amount of algae in the rock, however, is retained as a lithofacies because the rock is essentially composed of these organosedimentary structures. The stromatolitic dolostone includes both hemispherical and undulatory stromatolites (Plate 11, Figs. 1, 2 and 3). The algal

laminites are planar to wavy laminated and differentiated from the dolosiltites of member A by the presence of a bituminous film that is presumed to be algal in origin. Thin beds and lenses of dolosiltite, characteristic of member A, occur in member B.

Associated sedimentary structures include ripple marks, synaeresis cracks, desiccation polygons and intraclasts. Locally, lenses of gastropod dololutite occur with the stromatolites. The association of gastropods with the stromatolite beds possibly reflects the grazing habit of the gastropods on the algae. The association of gastropods and algae on Holocene tidal flats was documented by Garrett (1977). Other fauna consists of ostracods, horizontal burrow systems and the rare brachipod.

Discussion: Member B was probably deposited in the low intertidal zone. The absence of desiccation features in the interarea sediments between the stromatolites and the presence of desiccation polygons on the crests of the stromatolites suggests permanent wetting of the interareas and intermittent exposure of the stromatolites. Relief between the crests of the stromatolites and the interarea sediments is generally 0.3-0.5 metres with a maximum of 1.1 metres. This suggests that the tidal range, during formation of these hemispherical stromatolites, was low.

The synaeresis cracks in this member and other strata of this study are generally found in the rippled dolosiltites (Plate 6, Fig. 2). These dolosiltites occur as

continuous beds or as interarea sediments between hemispherical stromatolites. Synaeresis cracks can result from several mechanisms: 1) response to salinity variations (Burst, 1965; Donovan and Foster, 1972; Schwarz, 1975), 2) sediment loading (Plummer and Gostin, 1981) and 3) the dewatering of clays (White, 1961; Matter, 1967; Schwarz, 1975; Plummer and Gostin, 1981). In the strata of this study, the first mechanism is more probable and further supports the proposition of seasonal variations in the moisture levels and/or salinity of the tidal flat. Burst (1965) considered that the salinity changes due to seasonal variations in the surface run-off could initiate the development of synaeresis cracks. Synaeresis cracks in stromatolitic dolostones in the Lower Muschelkalk were attributed to salinity variations (Schwarz, 1975).

Recognition of the different tidal zones in members A and B is difficult since the criteria for delineating the zones are essentially the same (Fig. 6.1). The humid nature of the tidal flat also limits the demarcation of the zones as many of the diagnostic criteria are not present or limited in extent.

Member C

Description: Member C can be subdivided into three lithofacies: 1) bioclastic dolosiltite, 2) dolomitic limestone and 3) Type III mottled dolomitic limestone. These lithofacies represent the transition from the tidal flat to

the shallow subtidal zone. Unlike members A and B, member C is virtually quartz free. During deposition of member C, the Boothia Horst was a shallow, submergent feature and/or the site of active sedimentation as the facies belts shifted eastwards.

The bioclastic dolosiltite (lithofacies C1) contains ostracods and gastropods which commonly occur as lenses in the calcareous dolosiltites. Horizontal burrows such as *Gordia* (Plate 27, Fig. 2) and *Chondrites* (Plate 27, Fig. 1) are present. This lithfacies shows some evidence of mechanical deposition such as scour and ripple marks. Cross-laminae dominate and planar-laminae are essentially absent.

The dolomitic limestones (lithofacies C2) are gradational with lithofacies C1. Sedimentation appears to be cyclic and the contact between the micrite and the dolosiltite is sharp and shows evidence of scour. The dolosiltites are cross-laminated whereas the micrite is structureless. Megafossils include gastropods, ostracods, orthoconic nautiloids and rare brachiopods. Lithofacies C2 is delineated from the mottled dolomitic limestones (lithofacies C3) primarly on the basis of the lack of burrow structures and the presence of sedimentary structures in the dolosiltites.

Lithofacies C3 is a Type III mottled dolomitic biomicrite. The sedimentation rates were slow and the rates of bioturbation and early lithification of the micrite were

high. The genesis of these rock types was discussed in Chapter 5.

Discussion: The lithofacies of member C were deposited in the low intertidal to very shallow subtidal zone. Lithofacies C1 and C2 mark the transition into the subtidal zone and the basin margin-tidal flat interface. Lithofacies C3 indicates continual submergence in the shallow subtidal zone. The presence of erosion surfaces and hardgrounds in lithofacies C3 indicates the sediments were subaqueously rather than subaerially exposured.

Summary of the Cape Storm Formation

The strata of the Cape Storm Formation indicates that a tidal flat extended along the western margin of the Boothia Horst during the early Ludlovian. The east-west extent of the tidal flat is unknown. To the west was the M'Clintock Basin and to the east the Boothia Horst was probably a low-lying, emergent land mass.

The Boothia Horst may have acted as a barrier between the M'Clintock Basin to the west and the Prince Regent Basin to the east (Fig. 6.6). On Somerset Island, the Cape Storm Formation is 120-160 metres thick (Miall and Kerr, 1977) whereas on eastern Prince of Wales Island, the Cape Storm Formation has a maximum thickness of 97 metres. Lithologically, the Cape Storm Formation is similar on both islands. Thorsteinsson (1980) noted that the quartz content of the Cape Storm Formation is greatest on Somerset and

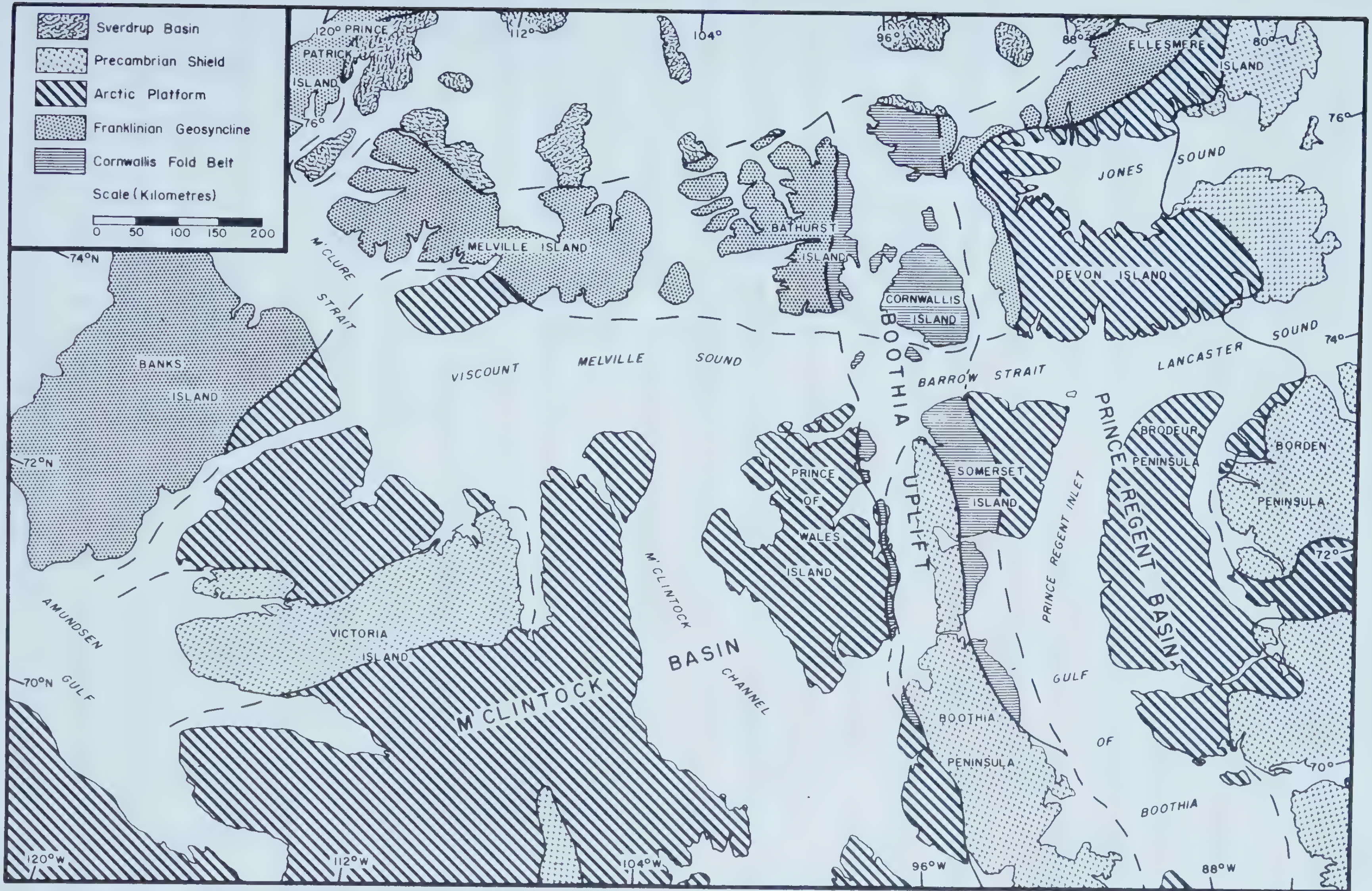


FIG. 6.6. Major structural provinces of the central Canadian Arctic Archipelago (from Tozer and Thorsteinsson, 1963; Miall and Kerr, 1980 and Thorsteinsson, 1980).

Prince of Wales islands. This further substantiates the role of the Boothia Horst in both controlling sedimentation and as a source area for the sediments.

The quartz silts are sub-angular which indicates a proximity to the source; however, the lack of feldspars, lithic fragments and micas would suggest that the Archean gneisses were not the source. The quartz grains are probably polycyclic and derived from the lower Paleozoic or Proterozoic strata such as the Lang River, Hunting and Aston formations. The quartz grains are generally monocrystalline and show uniform extinction under cross-polarized light. A similar argument can be presented for the origin of at least some of the detrital dolomites. The local variability in the quartz and detrital dolomite content may reflect local composition changes in the Boothia Horst.

Much of the strata of the Cape Storm Formation may be eolian sediments reworked by tidal processes. Eolian transportation of dolomite and quartz silts in the Persian Gulf was documented by Shinn (1973, 1983). Shinn (1973) noted that 65% of the eolian dust in the Persian Gulf are dolomite silts with quartz silts dominating the remainder. The dolosiltites of the Cape Storm Formation show a similar composition with the dolomite usually composing greater than 60% of the rock.

Event correlation in the Cape Storm Formation was not possible. Although a succession of lithofacies occurs, there is no time framework other than superposition implied by

this succession and no marker beds occur. A plot of the paleobathymetric curves (Fig. 6.7) shows no definite trends other than a transgressive sequence, ranging from supratidal at the base to shallow subtidal at the top of the formation. In approximately the middle of the Cape Storm Formation, there appears to be a minor regressive phase which separates the transgression into two cycles (Fig. 6.7). The lower cycle is more irregular and ranges from supratidal to low intertidal. The upper cycle marks a relatively smooth transgression from the high intertidal to the shallow subtidal.

The lack of correlation is not surprising when the nature of a tidal flat is considered. In an area with a small tidal range such as the M'Clintock Basin may have had, local physiographic features such as beach ridges, tidal pools or tidal channels can appear as as markedly different environments. However, the criteria used to distinguish the zones of a tidal flat are often shared by the different tidal zones and the physiographic features recognized on modern tidal flats are not always apparent in ancient rocks. One of the differences distinguishing the tidal zones is the periodicity of exposure; thus, the distinction between the different tidal zones may be contingent on the relative abundance of these criteria.

Section K is unique in that the upper two lithofacies (C2 and C3) of member C are not present and the transition into the Douro Formation is sharp (Fig. 6.2). The Cape Storm

FIG. 6.7. Paleobathymetric curves for the Cape Storm Formation on eastern Prince of Wales Island.

Formation is thin in this section (Fig. 6.3) but this does not appear to be anomalous to eastern Prince of Wales Island. Section K is also unique in that both the Douro and the Cape Storm formations are the thinest in this section. This may be related to syndepositional structure.

C. Douro Formation

Lithofacies

The relatively monotonous lithological and paleontological character of the Douro Formation on eastern Prince of Wales Island limits the delineation of regional internal stratigraphy in this formation. The Douro Formation is fossiliferous, generally being a sparse biomicrite or biopelmicrite. The fauna varies little in diversity throughout the formation but there is a tendency for a slight increase in faunal diversity in the upper half of the formation. Mottled dolomitic biomicrites are the dominate lithofacies having both lateral and vertical continuity (Fig. 6.8) whereas the other lithofacies tend to be localized and have little or no lateral continuity.

The second most common lithology is mottled dolomitic biopelmicrite. This lithology is best considered a subfacies of the mottled dolomitic biomicrite facies rather than as a true lithofacies because this rock type is not uniquely different from the mottled dolomitic biomicrites. Generally, the biopelmicrites occur as thin units that grade laterally

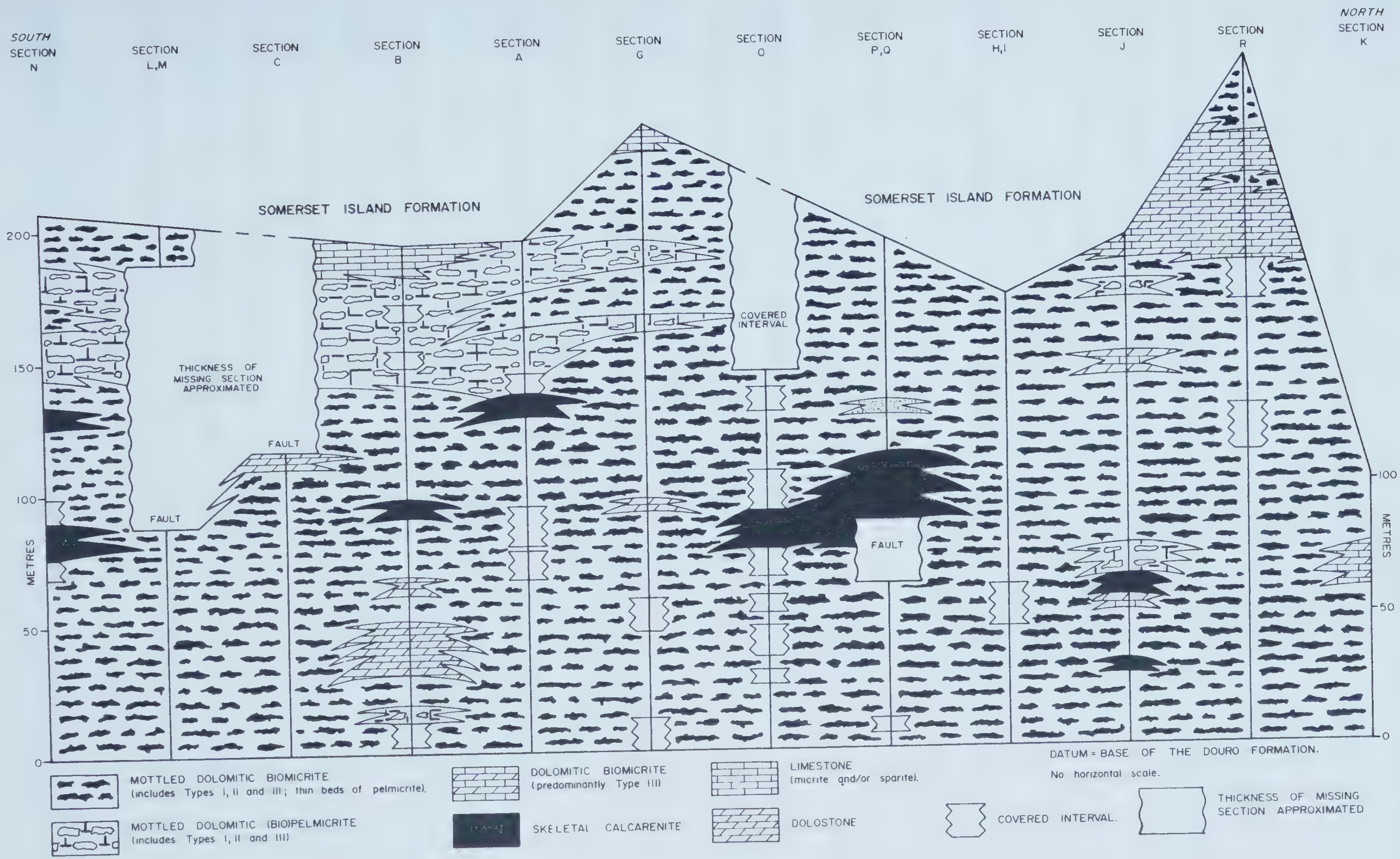


FIG. 6.8. Generalized regional lithofacies of the Douro Formation on eastern Prince of Wales Island.

into the biomicrites. Only in the southern area, Sections N to G, do the biopelmicrites form substantial units. In these sections, the biopelmicrites comprise approximately one-quarter of the Douro Formation (Fig. 6.8). The peloidal units are recognizable in the field but are best determined in thin section or polished slab (Plate 15, Fig. 3).

A second carbonate grain type (skeletal calcarenite) occurs in the Douro Formation. This lithology is differentiated from the pelmicrites in that the grains are of recognizable bioclastic origin and the rocks are grain supported. The calcarenites are usually isolated occurrences, forming field units or lenses in other units. The calcarenites were not traceable along strike. Typically, the calcarenites are cemented by sparry calcite and may be massive or rubbly weathering.

The skeletal calcarenites are shallow subtidal in origin, having formed under conditions of constant turbulence by wave or current action. They represent reworking of the biomicrites, winnowing out of the micrite and redeposition of the grains in new environments such as shoals or banks.

Dolostones are rare and occur only in the southern portion of study area (Fig. 6.8). The dolostones are late diagenetic in origin and commonly mimic the texture of the mottled dolomitic limestones (Plate 28, Figs. 1 and 2).

Detrital quartz is rare in the Douro Formation, generally occurring as isolated grains in the burrows. The

quartz content is low enough (1-2%) that it is only detectable in thin section. Section PQ is relatively anomalous in that it contains a 5.1 metre interval (units Q-19 and 20) of a mature, calcareous cemented, quartz arenite (Fig. 6.8, Appendix I). This is the only clastic unit in the Douro Formation and contains large-scale cross-stratification and *Megalomoidea* valves. This unit is traceable south along strike for approximately 0.5 km before it disappears beneath the overburden. In a stream valley, approximately 200 metres north of Section PQ, the quartz arenites are absent. The lack of intermixing, both laterally and vertically, with stratigraphically adjacent units further substantiates the uniqueness of this unit in the Douro Formation. The large-scale cross-stratification, the lack of faunal colonization and the lack of biogenetic structures indicates that this unit was a mobile and unstable substrate such as a sand wave deposited under turbulent conditions.

Micrite and sparite are rare; micrite occurs as field units whereas sparite occurs as fracture fill, cement and neomorphic calcite in other lithologies. Typically, micrite occurs as planar-laminated to massive units and are unfossiliferous.

No members or marker beds are recognized in the Douro Formation. The Douro Formation on eastern Prince of Wales Island is essentially a continuous sequence of mottled dolomitic biomicrites, whereas Narbonne (1981) and Narbonne

and Dixon (1982) recognized seven stratigraphic subdivisions of the Douro Formation on Cornwallis and Somerset islands. This subdivision was based on the argillaceous content and the presence of coral-sponge reefs. Neither the rubbly argillaceous limestones nor the reefs are present on eastern Prince of Wales Island.

Brachiopod zonation

No biofacies are recognized in the Douro Formation. The fossil assemblage may vary in population numbers but essentially the diversity remains consistant both laterally and vertically. The megafossils in the Douro Formation consists of trilobites, corals, stromatoporoids, brachiopods, echinoderms, gastropods, pelecypods, ostracods, dasycladacean algae, coralline algae and burrow structures. Brachiopods, in particular, the genus *Atrypoidea* dominates the fauna of the Douro Formation.

A zonation of *Atrypoidea* species is present in the strata of this study (Fig. 6.9). This genus is most abundant in the Douro Formation but also occurs in the Cape Storm and Somerset Island formations. In all three formations, there is a tendency for *Atrypoidea* to be associated with the mottled dolomitic limestones, suggesting a facies control on the distribution of this genus.

Approximately 26,000 brachipods were collected. Two genera dominate, *Atrypoidea* and *Protathyris*; the former being the most common by a ratio of approximately 10:1. Four

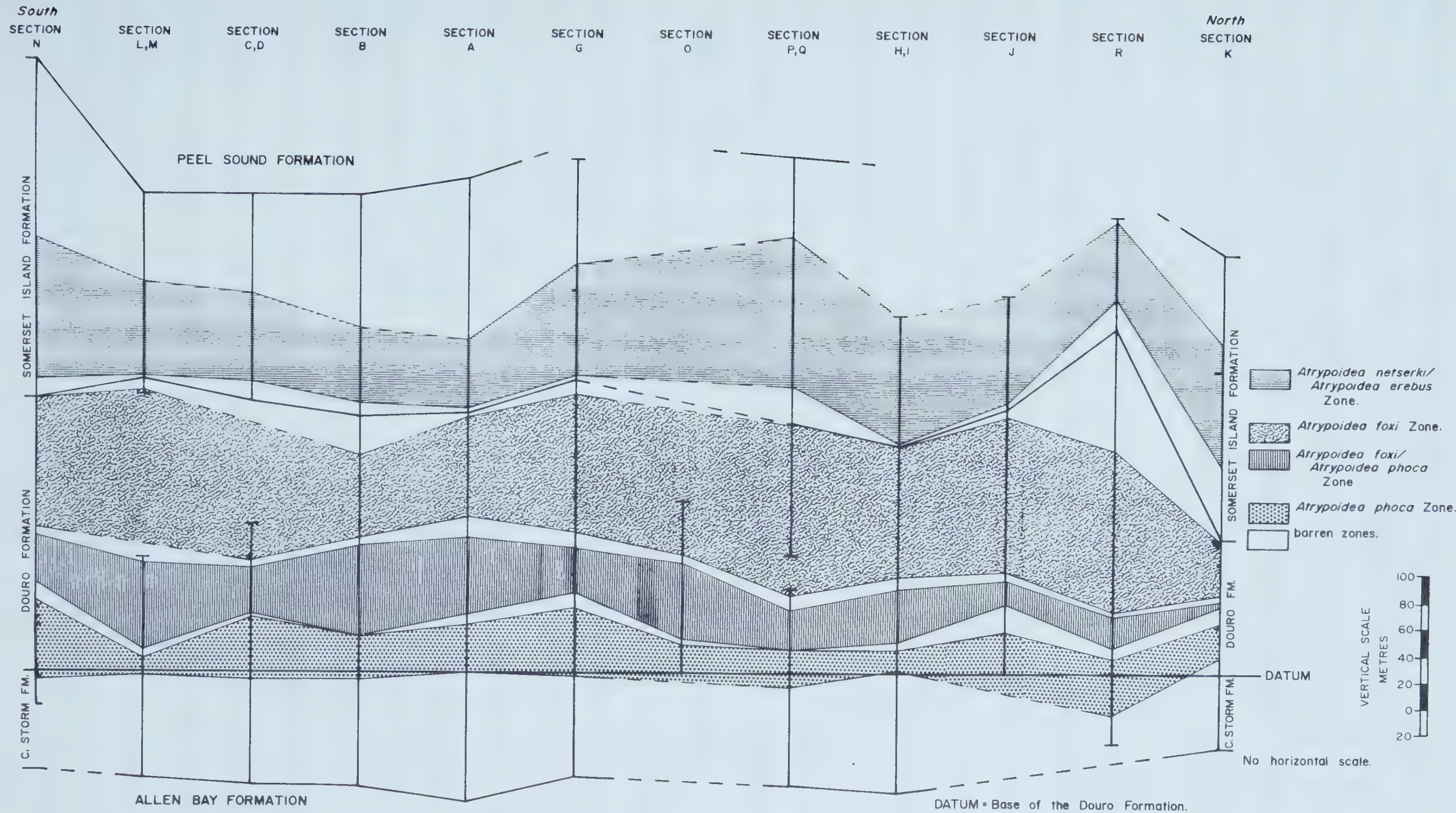


FIG. 6.9. *Atrypoidea* zonation on eastern Prince of Wales Island.

species of *Atrypoidea* occur and four zones are recognized (Fig. 6.9). Jones (1979, 1982) documented the same zonation throughout the south-central Arctic Archipelago. The *A. phoca/A. foxi* zone is a transitional zone between the two species whereas the *A. erebus/A. netserki* zone represents a mixing of the two species.

Jones (1981, p. 1539) attributed the *Atrypoidea* zonation to facies and evolutionary trends. As the zonation and the evidence of this study does not contradict Jones (1981) but rather supports his findings, his work will not be reiterated here. The stratigraphic range of *Atrypoidea* on eastern Prince of Wales Island shows a strong correlation with the stratigraphic distribution of the mottled dolomitic limestones. In the Cape Storm Formation, the introduction of *A. phoca* (Fig. 6.9) coincides with the introduction of Type III mottled dolomitic limestone (Fig. 6.2). This is most evident in Section R where both the *A. phoca* Zone and the mottled dolomitic limestones extend lower into the Cape Storm Formation than in other sections. In Section K, the mottled dolomitic limestones (lithofacies C3) are not present and the *A. phoca* Zone does not occur until the mottled dolomitic limestones of the Douro Formation.

No strong lithological control of the *Atrypoidea* occurs in the Douro Formation. Of note, however, is the depression of the *A. foxi* Zone in Sections B and R (Fig. 6.9). This is coincident with the shift from Types I and II mottled dolomitic limestones to Type III (Fig. 6.8) and further

suggests a facies control on the distribution of *Atrypoidea*. There is a slight correlation between the barren zones (Fig. 6.9) and minor regressive pulses in the paleobathymetric curves of the Douro Formation (Fig. 6.10). While this correlation is poorly defined, it may in part, account for the barren zones.

The *A. netserki/A. erebus* Zone (Fig. 6.9) is confined to the Somerset Island Formation and shows a strong correlation with the distribution of Type III mottled dolomitic limestones in this formation. This zone is unique relative to the other zones in that the *Atrypoidea* are commonly the only megafossils found in the stratum and the brachiopods occur in small pockets with large populations (Plate 15, Fig. 2). *A. phoca* and *A. foxi* occur with a higher diversity fauna and tend to be less numerous.

It should be noted, however, that the correlation between the *Atryoidea* and the mottled dolomitic limestones may be a function of the mottled dolomitic limestones being the dominant subtidal facies on eastern Prince of Wales Island. Elsewhere in the Arctic Archipelago, Jones (1979, 1982) has documented the occurrence of *Atrypoidea* in argillaceous limestones. This lithology is not present in the study area.

Section K is unique in that the Douro Formation is anomalously thin (Fig. 6.3). No evidence of faulting was found and this thin succession can be traced along strike for about approximately 10 kilometres to the north tip of

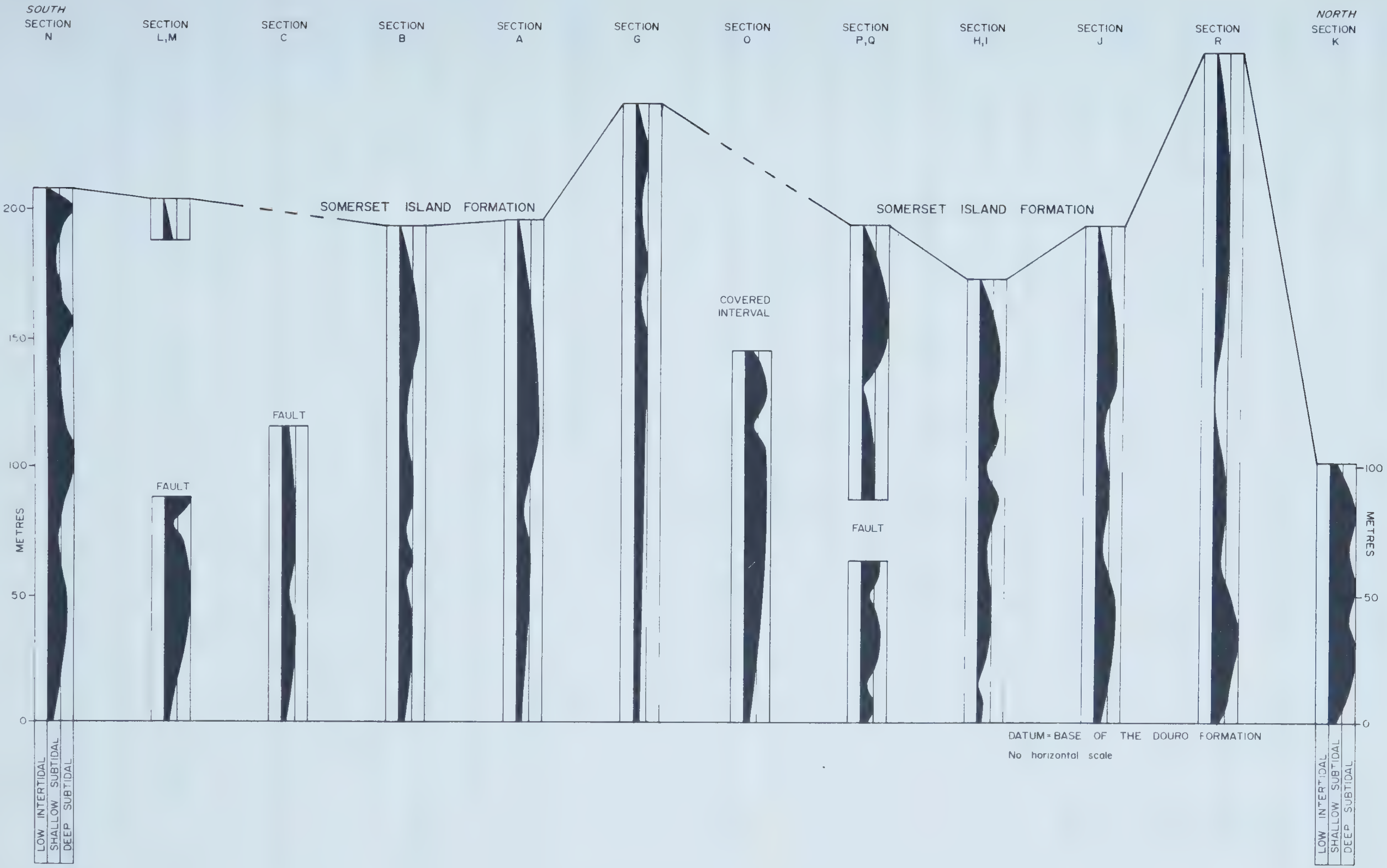


FIG. 6.10. Paleobathymetric curves for the Douro Formation on eastern Prince of Wales Island.

Prescott Island (Fig. 2.2). The only occurrences of digitate stromatoporoids and the rugose coral *Conophyllum* on eastern Prince of Wales Island occur in this section. These fossils also occur in the rubbly argillaceous limestones of the Douro Formation on Griffiths Island to the north of the study area.

Although the Douro Formation is reduced in thickness in this section, the presence of the *Atrypoidea* zonation (Fig. 6.9) further supports the contention that the thinning is related to syndepositional controls rather than postdepositional faulting. The extreme variation in thickness of the Douro Formation between Sections R and K (Fig. 6.3) is attributed to basement controls and the formation of several small fault-bounded sub-basins within the M'Clintock Basin.

Summary of the Douro Formation

The M'Clintock Basin on eastern Prince of Wales Island, during deposition of the Douro Formation, was a stable shallow platform. The Douro Formation represents the deepest stage of the transgressive-regressive cycle in the Late Silurian. Paleontological and lithological evidence indicates that the basin was generally within the effects of wave and/or current action.

The criteria for recognition of the shallow versus the deep subtidal zone in the strata of this study are similar (Fig. 6.1). Some of the problems in the recognition of these

two zones are: 1) basically the two have the same fauna, a marginal increase in faunal diversity occurs with increased water depth, 2) differentiation between shallow and deep is commonly based on the presence of sedimentary structures, winnowed grain-supported rock types and fragmented megafauna, 3) slight topographic changes on the depositional surface may cause locallized shallow conditions and regional trends may be absent and 4) local variations in the intensity, periodicity and bathymetry of waves or currents may lead to misinterpretation of the subtidal zones.

Paleobathymetric curves for the Douro Formation (Fig. 6.10) do not show any strong correlatable trends. There is, however, a weak cyclicity showing three deeping and shallowing cycles in the Douro Formation. These cycles are not strong enough to base a reliable correlation between sections. The paleobathymetric curves further supports the premise that most of the Douro Formation was deposited under shallow subtidal conditions near or above normal wave base.

The fauna is commonly fragmented and disrupted from life position. This includes large robust forms such as *Favosites* and stromatoporoids up to 40 cm in diameter. Commonly, the *Favosites* and stromatoporoids show evidence of several cycles of fragmentation and subsequent regeneration, indicating that the colonies were within at least intermittent wave or current action or storm wave base.

Lenses of skeletal calcarenite, usually composed of crinoidal debris, occur throughout the Douro Formation.

Intraclasts and sedimentary structures such as ripple marks (Plate 15, Fig. 1) are less common but also occur throughout the formation. The above features suggest deposition was predominantly within the effects of wave and/or current action.

The Douro Formation is unique relative to the other formations of this study in that it contains little to no quartz detritus. This is attributed to the Boothia Horst being a very shallow submergent or low-lying and areally restricted emergent feature. During the maximum transgression in the late Ludlovian, the Boothia Horst was possibly onlapped by sediments of the Douro Formation.

The composition of clasts in the lower Peel Sound Formation gives an indication of the composition of the Boothia Horst immediately after deposition of the Douro Formation. Miall (1969), during his work on the Peel Sound Formation, found that the lower conglomerates contained abundant clasts of Proterozoic strata and Archean crystalline rocks, common clasts of the Read Bay (=Douro) Formation in the basal conglomerate beds at some localities and few clasts of the Allen Bay Formation.

If the percentage of clast types is considered to be an index to the composition of the Boothia Horst, then the Douro Formation must have been limited in areal extent or have been a thin cover onlapping the Boothia Horst. The presence of clasts of the Douro Formation in the conglomerates demonstrates the the Douro Formation extended

eastward of its present-day distribution. The virtual absence of clasts derived from the Allen Bay Formation, supports the assumption that the Boothia Horst was essentially emergent until deposition of the Douro Formation. Brown *et al*. (1969) considered the Boothia Horst to have been shallow submergent feature or emergent as a few low-lying islands or a peninsula during deposition of the Allen Bay Formation.

The Douro Formation on eastern Prince of Wales Island is unique to the southern Arctic Lowlands in that the rubbly argillaceous limestones are not present. Narbonne and Dixon (1982) suggested the source of the terrigenous detritus in the rubbly limestones on Somerset Island was from the northeast. Trettin (1979) recognized Pearya and /or the Rens Fiord Uplift in the northwestern Arctic Archipelago as a source for the Cape Phillips Formation. Narbonne and Dixon (written communication to Jones, 1981) suggested the source for the argillaceous material in the Douro Formation may be the same as for the Cape Phillips Formation. The absence of the terrigenous material in the Douro Formation on eastern Prince of Wales Island further supports the proposition that the Boothia Horst was a barrier between the two basins during the middle to late Ludlovian.

D. Somerset Island Formation

Prior to this study, the Somerset Island Formation was only recognized on Somerset Island (Gibling, 1978; Miall and Gibling, 1978; Miall *et al.*, 1978). The Somerset Island Formation on Somerset Island, prior to 1978, was recognized as an unnamed transitional sequence between the Read Bay (=Douro) Formation and the Peel Sound Formation (Reinson *et al.*, 1976; Gibling and Narbonne, 1977; Jones and Dixon, 1977; Miall and Kerr, 1977).

For the purposes of discussion, the Somerset Island Formation on eastern Prince of Wales Island is divided into a lower part dominated by limestones, predominantly Type III mottled dolomitic biomicrites and an upper part of variable lithologies dominated by quartzose dolostone and quartz arenite/siltstone. These parts are lithologically similar to the two members of the Somerset Island Formation on Somerset Island Formation described by Miall and Gibling (1978) and Miall *et al.* (1978). However, this nomenclature will not be implemented for two reasons; one is the absence of the red siltstone which separates the two members on Somerset Island and two is the gradational contact between the two parts on eastern Prince of Wales Island.

The Somerset Island Formation shows abundant facies changes parallel to the Boothia Uplift. This is more pronounced in the upper part of the Somerset Island Formation than in the lower part of the formation (Fig. 6.11). The narrow outcrop belt on eastern Prince of Wales

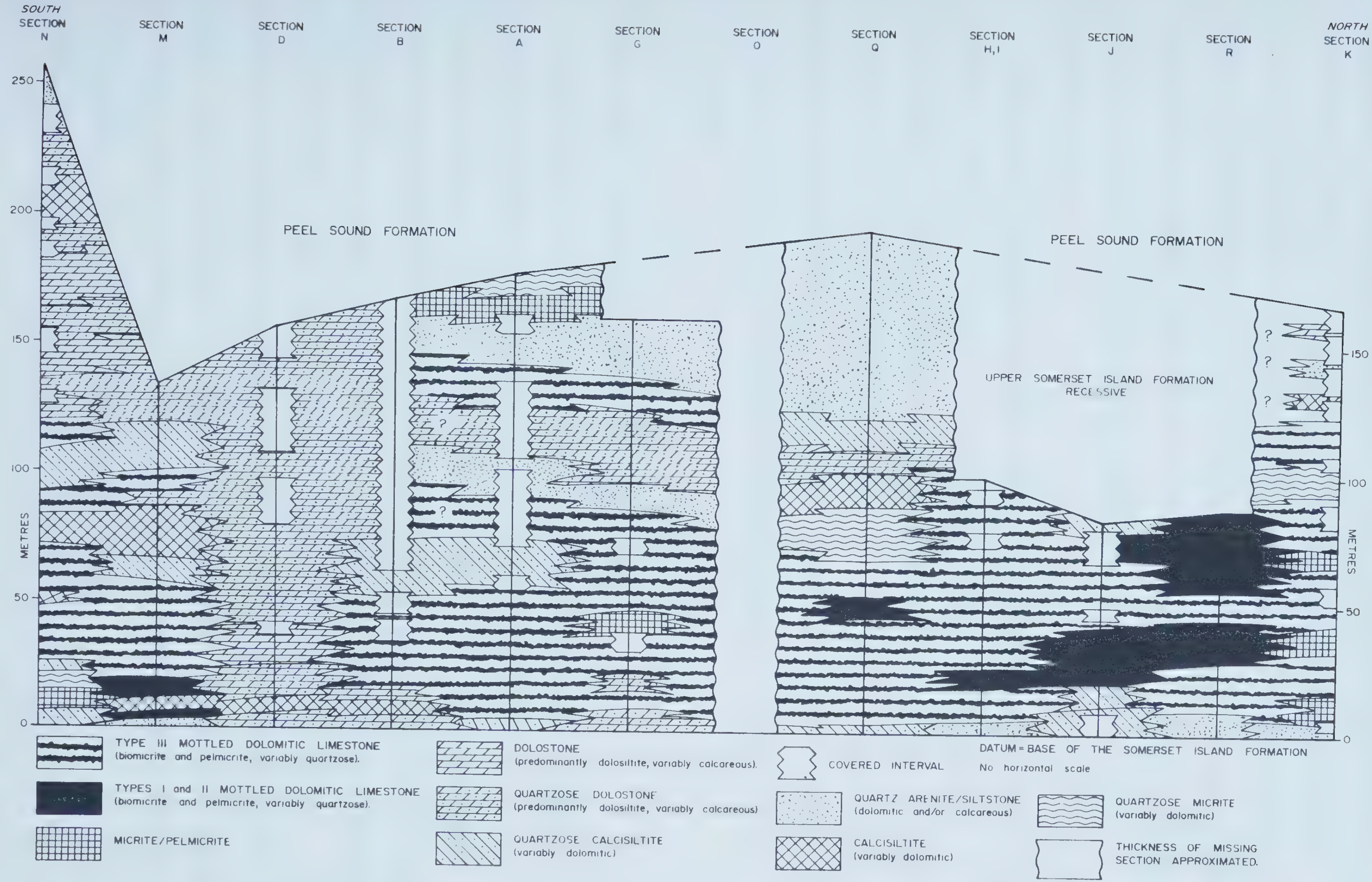


FIG. 6.11. Generalized regional lithofacies of the Somerset Island Formation on eastern Prince of Wales Island.

Island did not permit mapping of any strata perpendicular to the Boothia Uplift or to depositional strike. Miall and Gibling (1978) and Miall *et al.* (1978) noted that the two members of the Somerset Island Formation on Somerset Island interdigitate both parallel and perpendicular to the Boothia Uplift and that the two-fold subdivision is not always recognizable.

Several features of the Somerset Island Formation are apparent and distinguish it from the underlying Douro Formation: 1) the influx of quartz detritus, 2) the change in lithology and weathering pattern from the underlying Douro Formation, 3) the increase in quartz content and dolostone up-formation and 4) the decrease in fauna diversity.

Lower part, Somerset Island Formation

Description: The lower part of the Somerset Island Formation is dominated by Type III mottled dolomitic biomicrites with lesser amounts of Types I and II mottled dolomitic limestones, calcisiltite, micrite, quartz arenite and dolostone (Fig. 6.11). The lithofacies commonly contain all three end members (dolomite, calcite and quartz). Lateral variations are both depositional and diagenetic in origin. Sedimentary structures include desiccation polygons (Plate 6, Figs. 1 and 3), flat pebble conglomerates (Plate 5, Fig. 2), intraclast breccias (Plate 6, Fig. 4), synaeresis cracks (Plate 6, Fig. 2), ripple marks (Plate 5,

Fig. 3) and cross-laminae. The fauna includes brachiopods, gastropods, ostracods, corals, stromatoporoids and algal structures. Commonly, the fauna of a stratum is dominated by a single taxa such as brachiopods or corals but a mixed fauna also occurs.

The Type III mottled dolomitic biomicrites in the Somerset Island Formation differ from those in the Cape Storm or Douro formations by the presence of detrital quartz in both the micrite and the dolosiltite and the presence of desiccation features in the thin interbeds of calcisiltite and dolosiltite. The thin interbeds commonly show a scoured contact with the mottled dolomitic limestones. The fauna of the mottled dolomitic limestones suggests marine, subtidal conditions whereas the desiccation features indicate intermittent subaerial exposure. Subtidal rip-up clasts (Plate 6, Fig. 4) occur at the base of the formation and offer further evidence of early submarine lithification of the micrite layers and lumps.

The calcisiltites and dolosiltites in the lower part of the Somerset Island Formation generally show evidence of mechanical deposition such as scour and are, in part, detrital in origin. Burrow structures obscure much of the lamination in these units and they commonly appear massive in the field. These units also contain synaeresis cracks which may indicate seasonal variations in salinity as discussed with the Cape Storm Formation.

The dolostone sequence in Section D initially appears anomalous to the lower part of the formation. In this section, however, the extensive dolomitization is of diagenetic origin. This interpretation is based on the fauna present in the strata and the presence of recognizable textures in thin sections and polished slabs that show the dolomite to mimic previous limestone textures (Plate 28, Fig. 4).

Discussion: The Somerset Island Formation marks the dominance of the regressive phase in the late Ludlovian on eastern Prince of Wales Island and the return to tidal flat sedimentation. The constant influx of quartz detritus indicates that the Boothia Horst was a low-lying emergent land mass during deposition of the Somerset Island Formation. This was also suggested by Miall and Gibling (1978) and Miall *et al.* (1978). Abrupt lateral changes in the lithofacies reflects changes in the topographic and physiographic features of the depositional surface. This may, in part, be controlled by differential subsidence along contemporaneous faults that mark the margins of the sub-basins of the M'Clintock Basin on eastern Prince of Wales Island. As in the other formations of this study, the Somerset Island Formation lacks any correlatable marker beds.

The lower part of the formation dominated by strata deposited in the very shallow subtidal zone of a tidal flat. The occurrence of desiccation features in the thin interbeds

indicates intermittent subaerial exposure of the subtidal zone. Gebelein *et al.* (1980) observed subaerial exposure of the shallow subtidal platform on Andros Island during exceptionally low tides. This is perhaps a more logical explanation for the presence of these features than to postulate small tectonic pulses causing a shift of the subtidal-intertidal interface. The similarity of features shared by the low intertidal and shallow subtidal zones makes distinguishing the two difficult.

The lower part of the Somerset Island Formation on eastern Prince of Wales Island is approximately the same thickness (80-100 metres) as the lower member (98 metres) on Somerset Island of Miall and Gibling (1978) and Miall *et al.* (1978). The lower member on Somerset Island contains a higher incidence of desiccation features and of laminated dolostones and limestones (Miall and Gibling, 1978) than is found in the lower part on eastern Prince of Wales Island. Miall and Gibling (1978, p. 102, Fig. 10) recognized a cyclicity of laminated dolostones and limestones and of bioclastic limestones and dolostones in the lower member on Somerset Island. The lower part on eastern Prince of Wales Island is dominated by mottled dolomitic biomicrites and no cyclicity of the strata was apparent.

Miall and Gibling (1978, p. 101) considered the depositional environment of the Somerset Island Formation on Somerset Island to be a hypersaline, tidal mudflat. The only evidence to suggest hypersaline or at least intermittent

hypersaline conditions on eastern Prince of Wales Island is the presence of synaeresis cracks. The presence of a more diverse open marine fauna such as corals, stromatoporoids and coralline algae, the lack of abundant desiccation features and the absence of evaporitic minerals or pseudomorphs indicates that the lower part of the Somerset Island Formation was probably deposited under hyposaline conditions.

Upper part, Somerset Island Formation

Description: The upper part of the Somerset Island Formation is highly variable; quartzose dolostone and quartz arenite/siltstone dominate with lesser amounts of calcisiltite, mottled dolomitic limestone and micrite. The sequence contains numerous lateral and vertical facies variations. The upper part is similar to the upper member of the Somerset Island Formation on Somerset Island as described by Miall and Gibling (1978) and Miall *et al.* (1978) but lacks the pronounced red coloration in the siltstones and dolosiltites.

Small ripple marks with an amplitude of less than 1 cm and wavelengths of less than 7 cm (Plate 32, Figs. 1 and 2) are the dominate sedimentary structures. Intraclasts, desiccation polygons and synaeresis cracks are rarer. The fauna is dominated by ostracods and gastropods with the rare nautiloid, rhodolith and stromatoporoid which may be storm transported. The *Atrypoidea* zonation extends into the

limestones of the upper part of the formation.

Discussion: The upper part of the Somerset Island Formation was probably deposited on a tidal flat; a plot of the paleobathymetric curves shows most of the upper part to be intertidal in origin (Fig. 6.12). The lack of planar-laminae and the presence of burrow structures suggests that the strata are intertidal rather than supratidal in origin. The dolosiltites are predominantly detrital in origin (Plate 34, Figs. 1 to 4; Figs. 6.4 and 6.5). Dolarenites are more common in this strata than in the Cape Storm Formation.

Lateral facies variations from dolomitic quartz arenites and siltstones to quartzose dolostones reflect local variations in the composition of the tidal flats which may, in turn, reflect the composition of the source area. Both lithologies show the same style of current ripple marks suggesting similar depositional conditions occurred on both the carbonate and clastic dominated tidal flats.

The nature of the quartz grains in the upper part of the Somerset Island Formation changes from subrounded, monocrystalline grains (Plate 35, Figs. 1 and 2) to subangular, polycrystalline and stressed quartz grains (Plate 35, Fig. 3). This probably reflects the influence of the Archean crystalline rocks as a sediment source. Locally, the lower Paleozoic and Proterozoic strata must have been eroded and the crystalline rocks exposed. Miall (1969, 1970a, 1970b, 1983) and Miall and Gibling (1978) documented

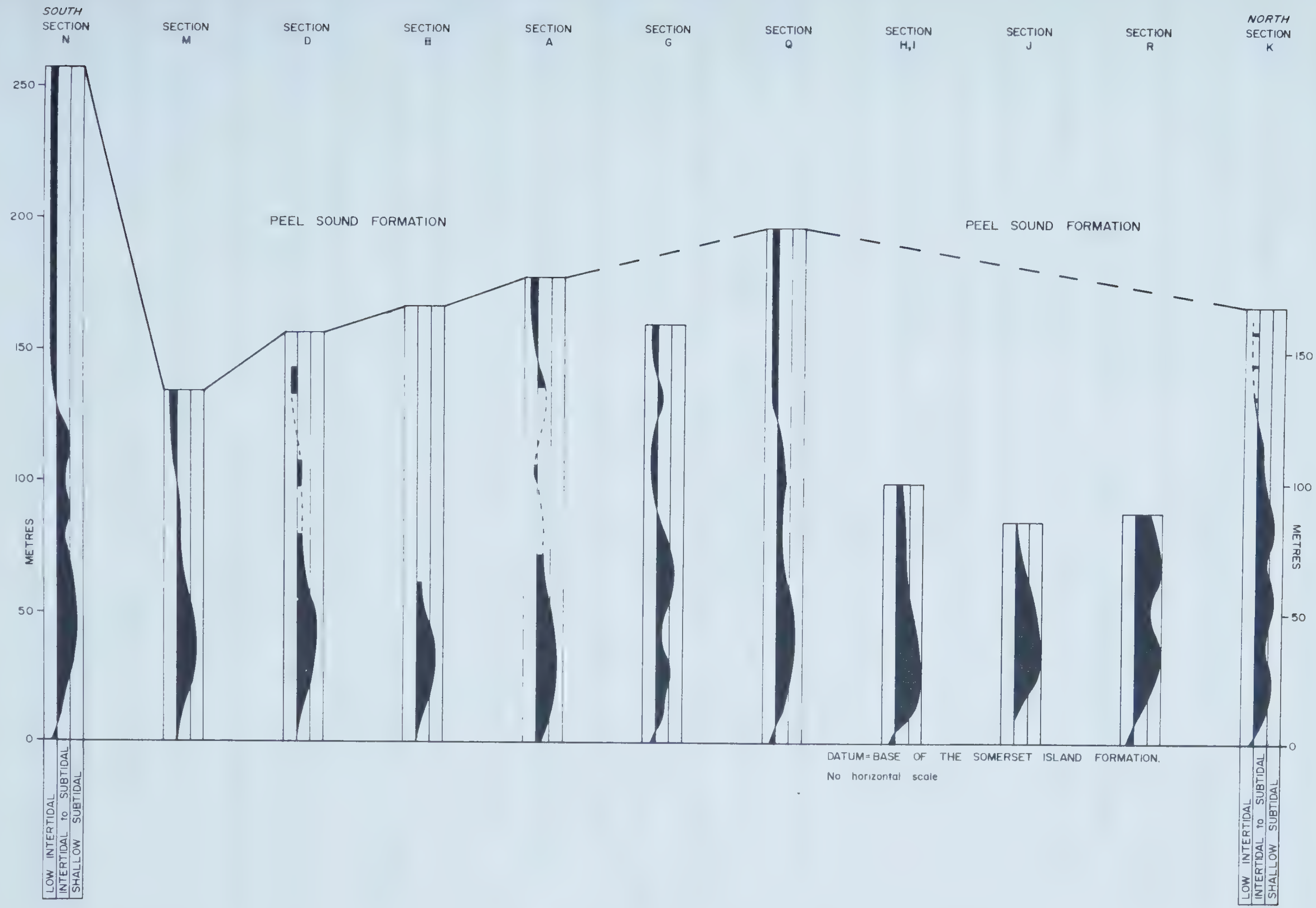


FIG. 6.12. Paleobathymetric curves for the Somerset Island Formation on eastern Prince of Wales Island.

the successive erosion of the lower Paleozoic cover during deposition of the clastic wedge in the Early Devonian. The nature of the quartz grains in the upper part indicates that the cover of sedimentary strata must have been breached as early as the late Ludlovian at some localities. This further substantiates the contention that the sedimentary cover of the Boothia Horst during the late ludlovian was thin and irregularly distributed about the Boothia Horst.

The clastic succession dominating the upper part of the Somerset Island Formation in Section PQ marks the introduction of a clastic dominated tidal flat. Miall and Gibling (1978) and Miall *et al.* (1978) documented deltaic deposits in the upper member of the Somerset Island Formation and the lower member of the Peel Sound Formation on Somerset and Prince of Wales islands, respectively. This occurrence of a clastic sequence in the Somerset Island Formation on eastern Prince of Wales Island indicates the locally diachronous nature of the clastic wedge or the diversity in the composition of the source rocks.

Amsbury (1962) documented fluvial transported dolomites which he considered to reflect the source area. At Strzelecki Harbour (Fig. 2.1), the tidal flat was dominated by detrital dolomites and the thickness of the upper part of the Somerset Island Formation in Section N (Fig. 6.11) indicates that a carbonate dominated tidal flat was sustained while a subaerially exposed alluvial/deltaic sequence developed immediately to the north. Miall (1969, p.

199) observed that the alluvial fans of the lower member of the Peel Sound Formation did not develop concomitantly along the western edge of the Boothia Horst.

There is little evidence for hypersaline conditions during deposition of the upper part of the Somerset Island Formation on eastern Prince of Wales Island. Miall and Gibling (1978), however, noted gypsum, halite casts and evaporite solution cavities in the dolosiltites and laminated sandstones of the upper member of the Somerset Island Formation on Somerset Island.

Regional correlation

The lower member of the Somerset Island Formation on Somerset Island is early Pridolian in age (Miall and Gibling, 1978; Miall *et al.*, 1978). This is in contrast to a late Ludlovian age of the lower part on eastern Prince of Wales Island. The conodont evidence is not conclusive as only one sample from the lower part of the Somerset Island Formation yielded conodonts. All the conodonts processed from the Upper Silurian strata on eastern Prince of Wales Island are in late Ludlovian *siluricus* and *latialata* zones (Uyeno, written communication, 1984; Appendix II). Thorsteinsson (1980, p. 2, Table 1), however, assigned the Somerset Island Formation on Somerset Island entirely to the *latialata* Zone.

Miall and Gibling (1978) and Miall *et al.* (1978) correlated the Somerset Island Formation on Somerset Island

with the lower member of the Peel Sound Formation on eastern Prince of Wales Island. They based this correlation on the assumption that the clastic wedges on both sides of the Boothia Horst would reflect the same tectonic episode. This assumption, however, was based on limited biostratigraphic evidence (Miall and Gibling, 1978).

A correlation between the Sandstone-Carbonate Facies of the upper member of the Peel Sound Formation on western Prince of Wales Island and the lower member of the Peel Sound Formation on eastern Prince of Wales Island was recognized by Miall (1969, p. 182). Gibling (1978, p. 260) and Miall and Gibling (1978, p. 115) also correlated the Somerset Island Formation on Somerset Island with the Sandstone-Carbonate Facies of the upper member of the Peel Sound Formation on western Prince of Wales Island (Miall, 1969) and the lower member of the Peel Sound Formation on eastern Prince of Wales Island. This correlation reflects the shift in facies belts in response to Pulse 2 of the Cornwallis Disturbance in the early Early Devonian (Kerr, 1977). The upper part of the Somerset Island Formation on eastern Prince of Wales Island possibly correlates with the Sandstone-Carbonate Facies of Miall (1969). This is a narrow, north-south trending belt of strata transitional between the redbeds to the east and the carbonates to the west.

The Carbonate Facies of the upper member of the Peel Sound Formation (Miall, 1969) on western Prince of Wales

Island is correlated with the lower part of the Somerset Island Formation on eastern Prince of Wales Island. Miall (1969, p. 187) described an open marine fauna with rare desiccation features in a limestone and dolostone sequence which is similar to the Type III mottled dolomitic limestones of the lower part of the Somerset Island Formation on eastern Prince of Wales Island. Mayr (1978, p. 24) and Thorsteinsson (1980, p. 15) assigned the Carbonate Facies of Miall (1969) to the Drake Bay Formation. On western Prince of Wales and Russell islands, they recognized a sequence of limestones and dolostones with some terrigeneous clastic detritus that overly the Douro Formation with a sharp and conformable contact. For the purposes of retaining uniform nomenclature about the Boothia Uplift, the Somerset Island Formation will be considered to have priority over the Drake Bay Formation.

Thorsteinsson (1980) considered the lower and upper members of the Peel Sound Formation on eastern Prince of Wales Island to be separated by a regional unconformity and used this to justify the change in nomenclature outside the Cornwallis Fold Belt on Prince of Wales Island. Miall (1983) argued that the Peel Sound Formation is a conformable sequence and only local angular unconformities exist between the lower and upper members.

Miall (1969, p. 183) considered the Carbonate Facies to be Lower and/or Middle Devonian in age whereas Thorsteinsson (1980) recognized a late Ludlovian age for some of the lower

strata of the Drake Bay Formation and a Pragian age for the upper strata. The lower age established by Thorsteinsson (1980) agrees with the late Ludlovian age determined by this study for the lower part of the Somerset Island Formation on eastern Prince of Wales Island. In response to Pulse 2 of the Cornwallis Disturbance, the eastern margin of the M'Clintock Basin shifted westward and the deposition of shallow subtidal carbonates was maintained into the Early to Middle Devonian.

E. Summary

The Upper Silurian strata on eastern Prince of Wales Island represents a transgressive-regressive cycle in which the dominant control appears to be the Boothia Horst (Fig. 6.6). Subsidence of the Boothia Horst during the early Ludlovian initiated the transgressive phase and subsequent deposition of the Cape Storm Formation. During deposition of the Douro Formation, the Boothia Horst was relatively stable. The regressive phase was initiated during deposition of the upper Douro Formation and may indicate the initiation of Pulse 2 of the Cornwallis Disturbance. The regression is most evident at approximately the Douro Formation-Somerset Island Formation boundary with the introduction of clastic detritus and subaerial exposure of the sediments; thus, marking the re-emergence of the Boothia Horst in the late Ludlovian. The effects of Pulse 2 are most apparent by the coarse clastics of the Peel Sound Formation.

The Boothia Horst played two important roles during the Ludlovian in the south-central Arctic Lowlands; one as a sediment source and two as a barrier between the M'Clintock and Prince Regent basins. The role as a sediment source is apparent by the presence of terrigenous dolomites and quartz in the Cape Storm and Somerset Island formations and the lack of such in the Douro Formation.

The differences in sedimentation across the Boothia Horst show its role as a structural barrier. On eastern Prince of Wales Island, the facies belts are condensed into a narrow facies belt whereas on Somerset Island Formation the facies are more extensive. The Ludlovian strata on eastern Prince of Wales and Somerset islands are similar in thickness and lithology. The pronounced asymmetry in the sedimentation does not become really pronounced until the coarse clastics of the Peel Sound Formation, at which time structural asymmetry is also most pronounced. This structural asymmetry is retained in the present-day structural of the central Arctic Archipelago.

Part of this asymmetry may be apparent rather than real. This is shown by the asymmetry of nomenclature about the Cornwallis Fold Belt and adjacent basins. Thorsteinsson (1980) applied a different nomenclature to Late Silurian and Early Devonian strata of the M'Clintock Basin than in the western margin of the Cornwallis Fold Belt or the Prince Regent Basin. This study has adopted the nomenclature from the Prince Regent Basin and applied it to both the

Cornwallis Fold Belt and the M'Clintock Basin on Prince of Wales Island. To date, work on the strata across the Boothia Horst has been done by different workers and added piecemeal to the literature with apparently little consultation between the workers. The resultant problems in the Upper Silurian stratigraphy and correlation about the Boothia Horst are obvious and need further clarification.

The recognition of regional trends in the Upper Silurian strata on eastern Prince of Wales Island is difficult due to the shallow water origin of the strata and the sensitivity of the sedimentary record to small variations on the depositional surface. Superimposed on the regional trends are the effects of differential subsidence of the basin margin. The subenvironments of the modern tidal flat such as recognized by Hardie and Garrett (1977) are not readily apparent in the strata of this study and only general tidal zones were recognized. Diagnositic criteria for the supratidal and high intertidal zones are similar as are the diagnostic criteria for the low intertidal and shallow subtidal zones.

This study area is unique because of the absence of rubbly argillaceous limestones found elsewhere in the Douro Formation of the Arctic Lowlands and in that it possibly represents a shallower sequence than is than documented elsewhere for the Douro Formation. The Ludlovian strata of this study represents a good example of a very shallow marine and transitional marine sequence of an ancient tidal

flat and shallow subtidal platform.

VII. STRUCTURE

A. Introduction

Two aspects of structural geology are considered in this study. One is the present-day structure and the effects faulting and folding have on accessing the stratigraphy. The strata of this study are part of the western margin of the Cornwallis Fold Belt (Fig. 6.6). The second is the role of tectonics in sedimentation during the late Ludlovian of the M'Clintock Basin. It is apparent that the Boothia Horst exerted considerable influence on the sedimentation patterns in the adjacent basins during this time.

Regional facies analysis has given rise to evidence that suggests the Boothia Horst became increasingly more unstable towards end of the Silurian; culminating in the Cornwallis Fold Belt. Although the Ludlovian appeared to be a period of relative tectonic stability, the eastern edge of the M'Clintock Basin was undergoing comtemporaneous deformation. Differential subsidence of syndepositional, fault-bounded sub-basins of the M'Clintock Basin is considered to be the mechanism largely responsible for the stratigraphic variations in the Cape Storm, the Douro and the Somerset Island formations (Fig. 6.3). These faults are possibly reactivated basement faults aligned perpendicular to the Boothia Horst.

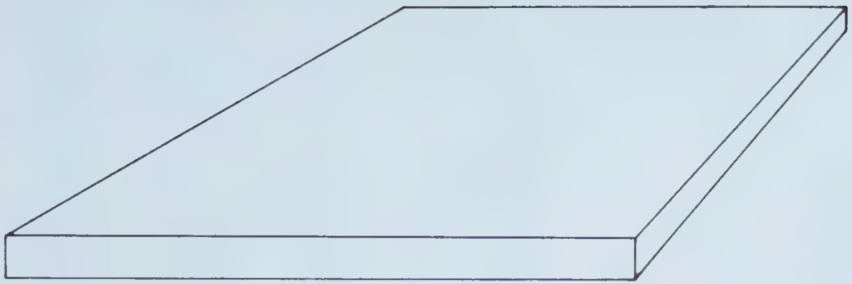
It is not the intent of this study to present a detailed structural analysis of the Cornwallis Fold Belt or

of the Boothia Horst. Rather the intent is to examine the implications of syndepositional and postdepositional structural events on the stratigraphy.

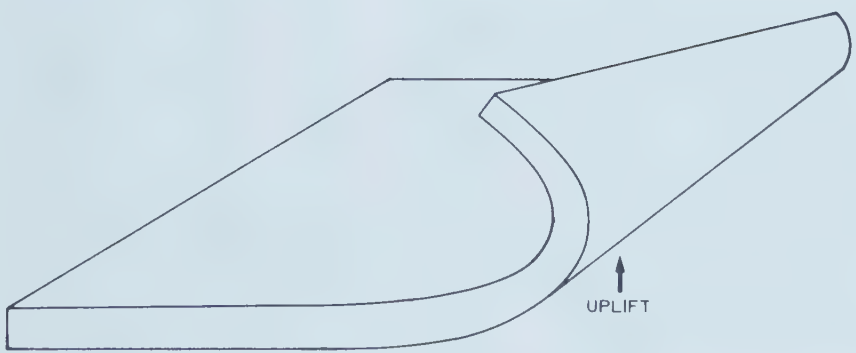
B. Folds

The structure of the western margin of the Cornwallis Fold Belt on eastern Prince of Wales Island can be generally described as that of a single asymmetrical syncline (Fig. 7.1). The western limb, which is gently dipping 2-10° to the west, is part of the essentially undeformed strata of the Arctic Platform. Lower Paleozoic strata of the western limb outcrop on western and central Prince of Wales Island (Fig. 7.2). The eastern limb is composed of lower Paleozoic and possibly some Proterozoic strata. Exposure is restricted to a narrow band of steeply dipping (greater than 60° west) to overturned (89-23° east) strata which has a maximum width of approximately 3 km but is generally less than 1.5 km wide (Figs. 2.1 and 2.2). The contact between the Cornwallis Fold Belt and the Arctic Platform is commonly sharp and faulted (Plate 36).

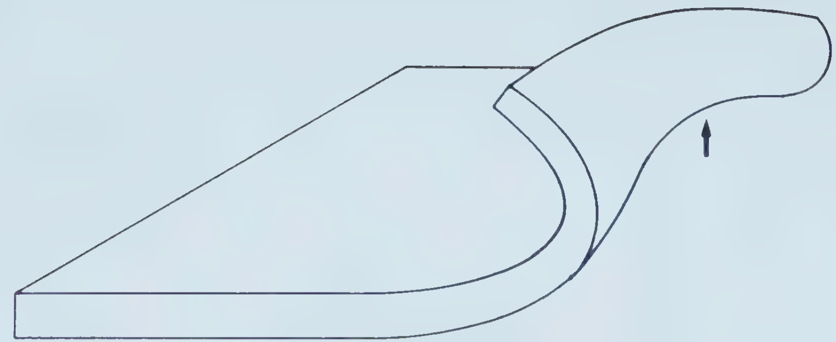
The present-day structure of the Cornwallis Fold Belt on eastern Prince of Wales Island ranges from an asymmetrical syncline to a monocline with dips decreasing to the west (Figs. 2.1 and 2.2). The variations in the nature of the fold can be attributed to several factors. One is the local deformation of the fold in response to increased local pressures (Fig. 7.1). This appears to be related to the



1. Continuous deposition in the M'Clintock Basin of the Lang River to Somerset Island formations.



2. Development of an asymmetrical syncline (Cornwallis Fold Belt) in response to uplift of the Boothia Horst in the early Early Devonian, Pulse 2 of Kerr (1977).



3. Subsequent or concomitant deformation of the asymmetrical syncline in response to local tectonic pressures during uplift.

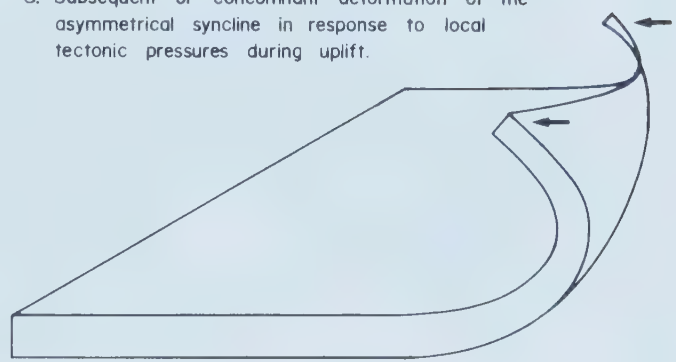
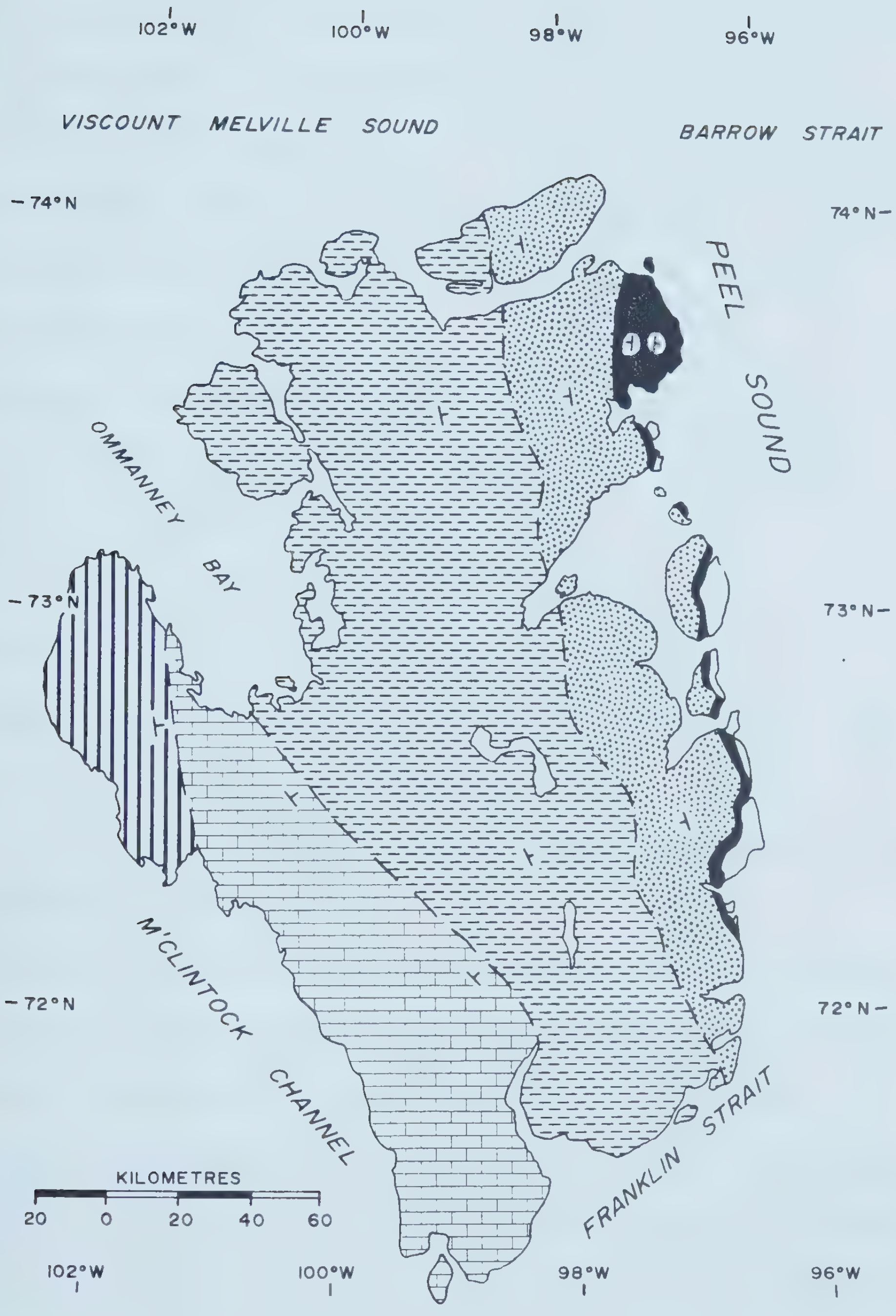


FIG. 7.1. Development of the "Flexure Bay" fold. The arrow indicates the area of maximum pressure exerted by the Boothia Horst.



Precambrian crystalline rocks and Proterozoic strata.
Pre-Douro Fm. Paleozoic strata.
Lower Paleozoic strata, western Cornwallis Fold Belt.
Douro Formation
Somerset Island Formation
Peel Sound Formation
Geological boundary (approximate) — — —
Regional strike and dip ⊥

FIG. 7.2. Geological map of Prince of Wales Island (after Christie *et al.*, 1971; Thorsteinsson, 1980; this study).

proximity of the Boothia Horst to the axis of the fold. Immediately south of the "Flexure Bay" fold, the strata change in attitude in response to "bending" around a Precambrian block (Fig. 2.1). A similar feature is evident at Savage Point where the strata are vertical to slightly overturned (90-88° west) and the full lower Paleozoic sequence is present (Fig. 2.1). Approximately 7 km south, the Douro and Somerset Island formations are steeply overturned (23° east) and abut against the Precambrian rocks (Fig. 2.1). The Allen Bay and Lang River formations of the eastern limb have been faulted out. Immediately to the west of Savage Point (Section HI), the KMG Decalta Young Bay F-62 well (72°41'25'' N, 96°49'30''W) penetrated a full lower Paleozoic sequence. Located on the western limb of the syncline, the strata penetrated by this well generally dip at less than 2° to the west.

The second variable is the level of erosion of the fold. Present-day erosional levels have exposed different structural levels of the fold and as a consequence, the strata show different attitudes. This may be the situation between Cape Brodie and Le Feurve Inlet where the monoclines are part of the western limb of the fold and represent a different structural level than the strata of the overturned limb.

Although the eastern limb of the syncline is overturned, the fold is not an overfold because both limbs dip in different directions (Billings, 1972, p. 50). Kerr

(1977) recognized both asymmetrical and overturned folds in the Cornwallis Fold Belt on eastern Prince of Wales Island but only the former were recognized in this study.

The fold is most evident at "Flexure Bay" (Plate 35, Fig. 4). This is the same structure referred to by Miall (1969, 1970b) as a syndepositional fold. Miall (1969, 1970b) based this interpretation on the clast composition in the conglomerates of the Peel Sound Formation and the presence of syntectonic conglomerates. The author cannot comment on the syndepositional nature of the fold since this strata was not included in this study. All the strata of this study and the strata immediately underlying and overlying are within the fold.

On the eastern limb of the syncline, small anticlinal folds with less than 1 metre amplitude occur in the lower part of the Somerset Island Formation. These are the only other folds observed on eastern Prince of Wales Island.

C. Faults

Faults both parallel and perpendicular to the Boothia Horst occur in Cornwallis Fold Belt on eastern Prince of Wales Island. Kerr and Christie (1965) and Kerr (1977) interpreted the western margin of the Boothia Horst as a steeply dipping reverse fault and considered the western margin to be predominantly a faulted contact whereas the eastern margin to be predominantly flexure and depositional contacts. Miall (1969), however, considered the western

margin of the Boothia Horst to be both flexed and faulted. A faulted contact between Lang River and the Precambrian rocks was recognized by Dixon (1973b). Christie *et al.* (1971) attributed all the contacts between the lower Paleozoic and Precambrian rocks to thrust faults.

Evidently the nature of the contact between the Cornwallis Fold Belt and the Boothia Horst is the subject of some controversy. The nature of this contact was not a prime concern of this study except where the Upper Silurian strata directly overlies the Precambrian rocks. In these areas, much of the lower Paleozoic strata is missing and the contact is faulted. When possible this contact was traced in the field but the contact is generally recessive and rarely visible. In areas where a more complete section of lower Paleozoic strata occurs, faulting was not evident but the fault contacts recognized by Dixon (1973b) are assumed to be present. The style of faulting is unknown and the contact is mapped as a geological boundary (Figs. 2.1 and 2.2).

Where the faulting is evident, it is assumed to be steeply dipping reverse faults as described by Kerr and Christie (1965) and Kerr (1977). Kerr (1977, p. 1391) considered the faults along eastern Prince of Wales Island to change with structural level, being a near vertical fault in the south and a high angle reverse fault northwards. Prucha *et al.* (1965) also noted that reverse faults can change in attitude, depending on the structural level observed. The attitude of the strata in the fold changes

with the structural level observed; therefore, it is likely that the nature of the reverse faults also change with the structural level observed along the western margin of the Boothia Horst. The faults possibly range from vertical faults to low angle reverse (thrust) faults.

The Cornwallis Fold Belt is segmented by faults aligned perpendicular to the Boothia Horst and that extend from the Precambrian rocks through to the Peel Sound Formation or are entirely within the lower Paleozoic strata (Figs. 2.1 and 2.2). Some of these faults are probably reactivated basement faults that were active either throughout the Late Silurian or reactivated during development of the Cornwallis Fold Belt in the latest Silurian to Early Devonian times. Reactivated basement faults, parallel to the north-south structural trend of the Boothia Horst, were documented by Kerr (1977).

Small west-dipping reverse faults occur in the eastern limb of the syncline. Typically, these faults dip to the southwest or west at 22-24° and have a maximum displacement of 6 metres. Generally the displacement is less than 1 metre and a series of these small faults form small imbricate structures. Slickensides and the offset of the strata indicates that the upper block moved eastward relative to the lower block.

Jointing is common and is most apparent on fresh outcrops of the Douro Formation. One set is parallel to or along bedding planes while the other set is oblique to

bedding. The presence of calcite slickensides on the joint surfaces indicates that some movement has occurred along the joints. The disposition of the joints relative to the fold was not documented but it is assumed the joints resulted from the same compressive forces as produced the fold. Rare occurrences of en échelon calcite veins occur oblique to bedding and suggest some extensional fracturing of the rocks. The development of the Peel Sound is attributed to extensional block faulting along reactivated structures in the Boothia Horst during the Cretaceous-Tertiary Eurekan Rifting Episode (Kerr, 1977). The calcite veining may be related to this extensional event.

D. Sub-basins of the M'Clintock Basin

There are stratigraphic anomalies in the Upper Silurian strata on eastern Prince of Wales Island that do not appear to be related to postdepositional structure and that may be explained by syndepositional tectonics. Previous discussions have demonstrated that the Boothia Horst was tectonically unstable during the Late Silurian and that this instability exerted considerable influence on the sedimentation in the adjacent basins. It is proposed that the eastern margin of the M'Clintock Basin was segmented by reactivated basement (contemporaneous) faults into five sub-basins (Fig. 7.3).

The north and south margins of the sub-basins are controlled by faults perpendicular to the basin margin but the control of faulting parallel to the basin margin is not

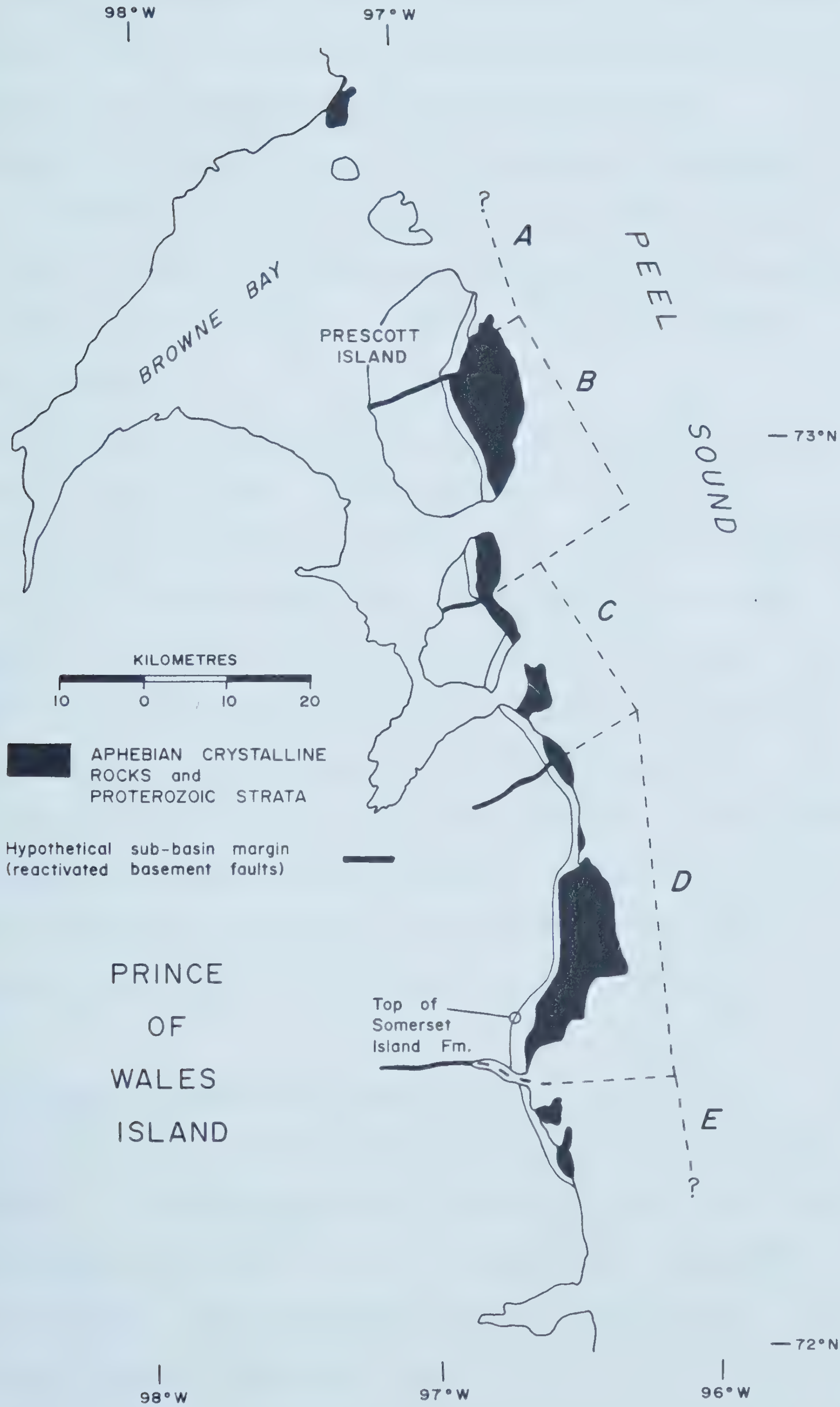


FIG. 7.3. Suggested sub-basins of the M'Clintock Basin on eastern Prince of Wales Island.

known. The mechanism of fault reactivation is also unknown but due to the slow rates of sedimentation during the late Ludlovian, it is assumed that faulting proceeded independently of deposition. The importance of basement related contemporaneous faults in the subsidence of small basins and in which the movement originated in the basement independent of depositional controls was recognized by Shelton (1968).

Although five sub-basins are recognized on eastern Prince of Wales Island, the discussion will centre on sub-basins A and B (Figs. 7.3 and 7.4). The most pronounced differences occur between these two sub-basins and the proposal of fault controlled sub-basins can be best documented on Prescott Island (Fig. 7.4). Two important criteria for the recognition of the sub-basins are stratigraphic variations in the Douro Formation that cannot be explained by repetition or omission of strata by faulting and the presence of alluvial fans overlying the basin sediments. The role of the former will be discussed first.

Stratigraphic evidence for sub-basins

Stratigraphic evidence for the sub-basins includes: 1) absence of postdepositional faulting in Sections K and R, 2) significant variation in the thickness of the Douro Formation over a short distance and 3) lithofacies variations between Sections K and R.

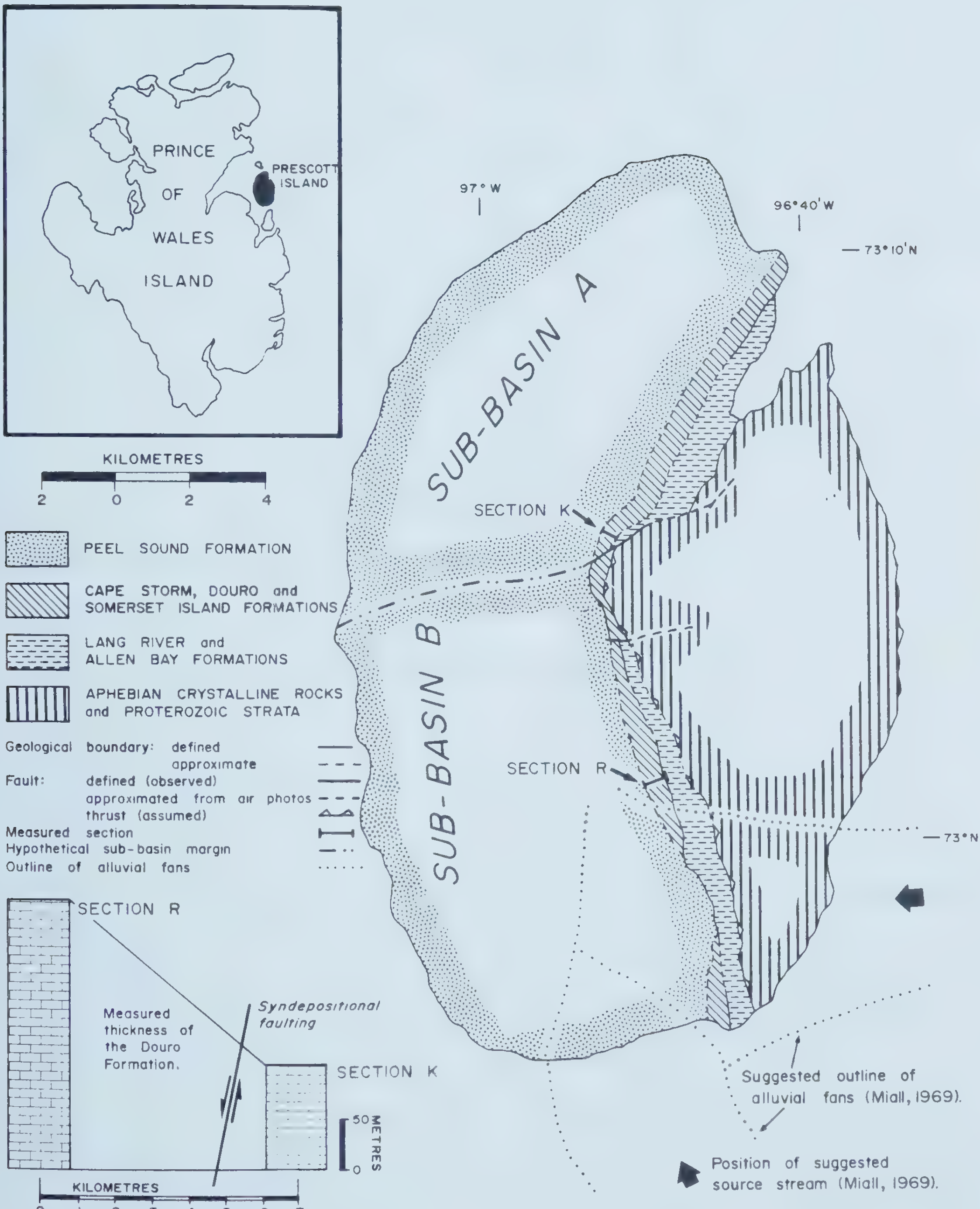


FIG. 7.4. Sub-basins of the Upper Silurian on Prescott Island.

In neither Section K nor Section R was there evidence to suggest repetition or omission of strata due to postdepositional faulting. Between these two sections, a distance of approximately 7 km, the thickness of the Douro Formation varies by 160 metres (Fig. 7.4). The proposed model for the sub-basins must address two anomalies in the Douro Formation. One is the abnormally thick succession (262 metres) in Section R and the other is the abnormally thin succession (101.5 metres) in Section K (Fig. 7.4). In other localities (Sections LM and CD), the thinning of the Douro Formation due to faulting was readily apparent. Section K is also unique in that the Cape Storm Formation is thin (56.5 metres) relative to the partial section (52.6 metres) measured in Section R (Fig. 7.4).

The homogeneous nature of the mottled dolomitic limestones and the lack of marker beds in the Douro Formation hindered the recognition of faults in this strata. Should any major repetitions or omissions of strata occur, it should be reflected in the distribution of the brachiopod genus *Atrypoidea*. The *Atrypoidea* show a well-defined zonation throughout the Upper Silurian strata on eastern Prince of Wales Island (Fig. 6.9). In Section K, the zonation is complete and shows a consistant reduction in zone thickness corresponding to a reduction in thickness of the Douro Formation. In Section R, the zones are relatively consistant in thickness with those found in other sections (Fig. 6.9). The only significant variation is the presence

of a thick barren zone at the top of the section which was shown to be facies related (Chapter 6).

The substantial differences in thickness of the Douro Formation on eastern Prince of Wales Island are attributed to differential subsidence of the sub-basins due to syndepositional faulting along reactivated basement faults (Fig. 7.5). Interpretation of the timing of these movements is subjective and based primarly on stratigraphic trends between and within sections. Movement along these faults possibly initiated during deposition of the upper member of the Cape Storm Formation and may not have been active in all the sub-basins at this time. The nature of tidal flat sedimentation and the lateral variability of these deposits does not allow for recognition of fault movement, if any occurred, below the upper member of the Cape Storm Formation. Differential subsidence of the sub-basins (Fig. 7.5) continued throughout the middle to late Ludlovian with the differences in the basins becoming more pronounced as deposition continued. No criteria was recognized to indicate whether the relative rates of subsidence varied with time.

In Sections K and R, the recessive nature of the Cape Storm Formation prevented correlation between the two sections. The differences in the upper member suggests that the fault was active during deposition of the upper member of the Cape Storm Formation. The stratigraphic position of the dolostones in the Cape Storm Formation in Section K is in approximately the same stratigraphic position as the

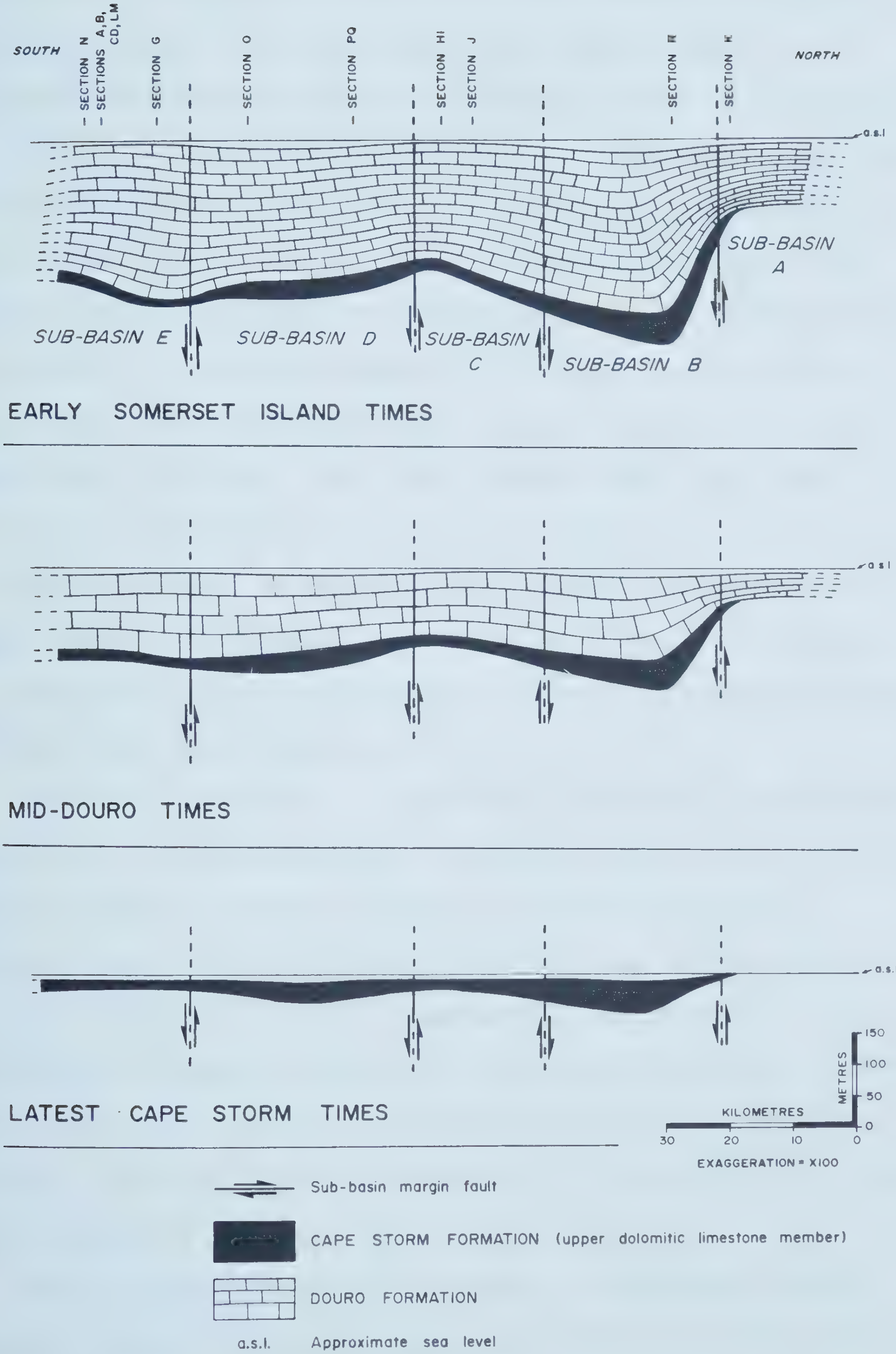


FIG. 7.5. Evolution of sub-basins of the M'Clintock Basin during the Ludlovian on eastern Prince of Wales Island.

dolostones (lithofacies C1 of the upper member) in other sections (Fig. 6.2), suggesting that they may be lithologically correlatable. Subsidence during this time may have been relatively uniform between the sub-basins and the faults may not have been active except between sub-basins A and B. Section R contains a thick dolomitic biomicrite and Type III mottled dolomitic biomicrite interval (lithofacies C2 and C3) in the upper member of the Cape Storm Formation (Fig. 6.2). This indicates that subtidal deposition of the upper member may have initiated and have been sustained earlier in sub-basin B than in other sub-basins. If the reduced succession in Section K is assumed to represent a slower rate of subsidence in sub-basin A than in sub-basin B, the chronological equivalent of this dolostone may be the mottled dolomitic limestone.

The Douro Formation in Section R is unique in two ways; one is the thickness and two is the presence of approximately 50 metres of Type III mottled dolomitic biomicrite at the top of the formation (Fig. 6.8). This is the only section in which a substantial thickness of Type III mottled dolomitic limestones occurs in the Douro Formation. While it may be possible to include this strata with the Somerset Island Formation, it is included with the Douro Formation since it lacks quartz detritus.

Should these 50 metres of strata be assigned to the Somerset Island Formation rather than the Douro Formation, the thickness of the Douro Formation in Section R (262

metres) would approximate the regional thickness of 200 metres. However, the lower part of the Somerset Island Formation in this section would be anomalously thick and suggest that subtidal deposition persisted in this sub-basin while intertidal deposition occurred elsewhere. Regardless to which formation this strata is assigned, it suggests the same conclusion; subtidal conditions were sustained in this sub-basin while the depositional conditions outside of this basin were variable. This shows some of the problems in the correlation of the strata of this study and the assignment of strata to formations.

Section R is also unique in that the lower part of the Somerset Island Formation contains substantial thicknesses of Type I and II mottled dolomitic biomicrite (Fig. 6.11). These rocks probably represent deeper subtidal conditions than Type III mottled dolomitic limestones and further suggests that the subsidence was greater in sub-basin B than in the other sub-basins. Section K contains subtle paleontological differences from other sections on eastern Prince of Wales Island. Digitate stromatoporoids and a solitary rugose coral (*Conophyllum*) with flanges of up to 10 cm occur only in the Douro Formation of this section.

Alluvial fans and sub-basins

Miall (1969) recognized a series of Devonian alluvial fans on eastern Prince of Wales Island (Fig. 7.6). In the areas where the two studies geographically overlap, there is

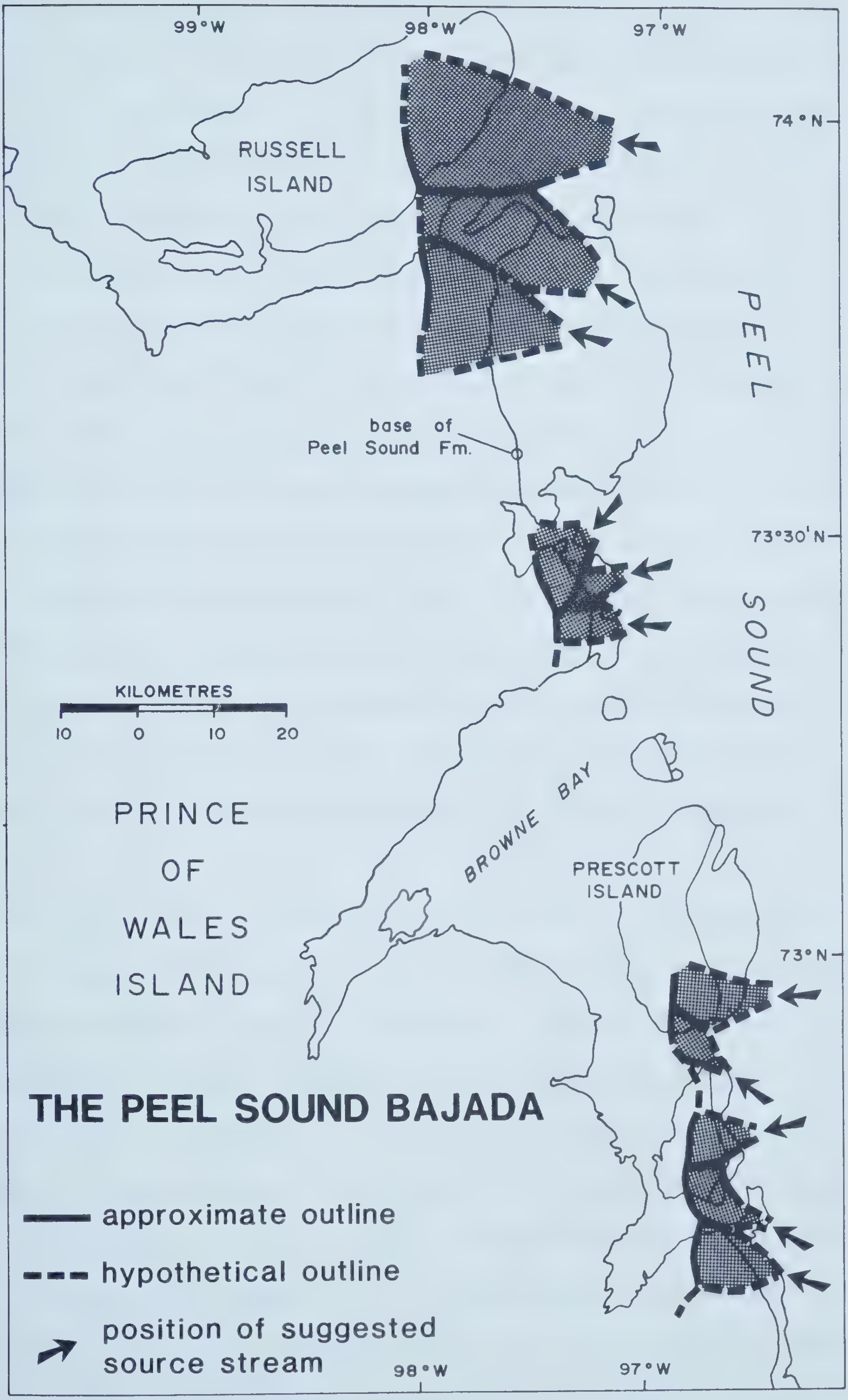


FIG. 7.6. Suggested outline of eleven alluvial fans - the Peel Sound Bajada (from Miall, 1969, p.86, Fig. 30).

a correlation between the boundaries of the alluvial fans (Fig. 7.6) and the margins of the Late Silurian sub-basins (Fig. 7.4). Therefore, it is proposed that the boundaries of the alluvial fans correspond with basement faults and that these faults were active into the Early Devonian.

The association of alluvial fans with tectonically active areas has been documented by Bryhni and Skjerlie (1975), Steel and Wilson (1975) and Steel (1976). Steel and Wilson (1975) recognized the role of faulting in controlling fan sedimentation and the geographic distribution of fans. They considered the coarsening upwards sequences in the fans as an indication of increased fault intensity. Steel (1976, p. 207) documented Devonian alluvial fans in Norway that are fault bounded along three margins and show faulting and folding of the alluvial fans. Folding of fan deposits by tectonic uplift was also documented by Bhyrni and Skjerlie (1975).

The Norwegian examples are analogous to the Devonian alluvial fans described by Miall (1969, 1970b). Miall (1969, 1970b) documented similar coarsening upwards cycles in the Peel Sound Formation, syntectonic conglomerates and syndepositional folding of the alluvial fans on eastern Prince of Wales Island. Two phases of conglomerate wedges were recognized by Miall (1969, 1970b) and Miall and Gibling (1978); an earlier phase of fan development which was later scavenged as a source material for the second phase. Two phases of alluvial fan development in the Permo-Triassic of

Scotland were documented by Steel and Wilson (1975).

A further consideration in this argument is the presence of alluvial fans on southern Prescott Island but not on northern Prince of Wales Island (Fig. 7.6). In Section K, the basal Peel Sound Formation is a red sublithic arenite which may represent deposits of an alluvial plain rather than a fan. The Peel Sound Formation was not present in Section R but traverses during 1982 and 1983 showed that the Peel Sound Formation on southern Prescott Island consists of sandstones and conglomerates. Steel (1976) documented similar differences in the coarseness of sediments in Devonian fault-bounded basins of Norway and attributed these differences to tectonic contexts during development of the basins. This difference is the basal Peel Sound Formation further supports the proposition of tectonic sub-basins.

Although Miall (1969) did not suggest fault bounded margins for the alluvial fans of the Peel Sound Formation, it is apparent that there is justification for the proposing fault margins to the alluvial fans. This is supported by comparison with other ancient alluvial fans and criteria present in the underlying strata.

The other sub-basins are not as well defined but show some or all of the features discussed above. Sub-basins A and B (Fig. 7.4) show the most pronounced differences and therefore, were the sub-basins used to explain the model. There is a possibility that more than five sub-basins

existed but the criteria are not well-defined.

While the evidence for the development of syndepositional sub-basins and differential subsidence is relatively strong, alternatives must be considered. The most obvious is the role of postdepositional faulting and while some criteria have been presented to discount this, the possibility of some influence by postdepositional faulting cannot be totally disregarded. In other sections, the faults were obvious and the amount of missing strata approximated (Fig. 6.8). It is possible that several faults are present in Sections K and R but were not readily apparent. A series of faults, spaced throughout the sequence and each with a displacement of a few tens of metres could explain why the *Atrypoidea* zonation is complete but reduced or increased in thickness in the respective sections. These faults would have to be conveniently placed to retain the zonation. This style of faulting, however, would be inconsistent with the faulting observed in other sections.

Another problem is the rate of sedimentation. Most of the argument for syndepositional faulting occurs in the mottled dolomitic limestones. As proposed in the model for the genesis of these rock types (Figs. 5.1 and 5.3), the rates of formation of these rocks are assumed to be relatively uniform both spatially and temporally. It is apparent that this model needs some revision if the thickness differences between the sub-basins are purely depositional. Locally, there must be variability in the

rates of sedimentation and hence the rates of bioturbation and lithification if this rock type is to be maintained. Alternatively there must be time-stratigraphic differences between the basins such that the thickness differences may be attributed to duration of deposition. The lack of marker beds and/or of time indicators in this strata hinder recognition of these two factors but it is likely that both functioned concomitantly to cause the differences between the basins.

E. Summary

The sedimentological evidence and the presence of the Cornwallis Fold Belt indicate that the eastern margin of the M'Clintock Basin was tectonically unstable throughout the Late Silurian and Early Devonian. The tectonic activity was relatively continuous but variable in intensity during this time. Oscillations of the Boothia Horst are apparent by the changes in sediment type during the transgressive-regressive cycle of the Late Silurian. The major tectonic events occurred in the Early Devonian (Pulse 2 of the Cornwallis Disturbance of Kerr, 1977) through to the Eurekan Rifting Episode in the Cretaceous-Tertiary. It was these tectonic events and the subsequent erosion of the uplifted strata that produced the present-day structure.

Tectonic activity during the Ludlovian was more subtle than in subsequent periods. Superimposed on the transgressive-regressive cycle are the effects of

differential subsidence of the sub-basins. The faults bounding these sub-basins were active into the Devonian and were zones of structural weakness during the Cornwallis Disturbance. They can be recognized in the present-day structure but are only apparent in the strata of this study by detailed stratigraphy and petrology.

The model proposed for the sub-basins appears to satisfy most of the differences in the stratigraphic sequence observed in measured sections and on traverses. The alluvial fans of the Peel Sound Formation were a useful source of support criteria for this model. While this model may not totally satisfy all the problems in the stratigraphy of the study area, it goes a long way to explaining some of the stratigraphic anomalies encountered in a basin margin adjacent to a tectonic uplift.

VIII. SUMMARY AND CONCLUSIONS

Three main contributions of this study to the geology of the south-central Arctic Archipelago are; the paleoecology of algae, the role of biogenetic processes in the formation of mottled dolomitic limestones and the influence of the Boothia Horst on sedimentation during the late Ludlovian.

Two groups of algae, the porostromates and the rhodophycophytes, occur in abundance in the strata of this study. The spongiostromates are locally abundant and have been documented throughout the Arctic Lowlands. Prior to this study, no systematic study of algae in the Arctic Archipelago was undertaken; algal structures were recognized but usually only mentioned. This is not to say that this study resolved the role and contribution of algae in/to the strata but rather that it is a groundwork for successive studies. Perhaps the most significant aspect of the algae was the recognition of the rhodoliths throughout a stratigraphic sequence and the adaptation of *Solenopora* to a spectrum of carbonate environments. Few studies document the diversity of algal structures and of their paleoenvironments as are recognized in the strata of this study.

All of the genera of porostromate algae occurring in this strata have been documented elsewhere in the Arctic Lowlands; usually as fragments and rarely as independent algal structures. It is apparent from comparative studies of the external morphology of spongiostromate, porostromate and coralline algal structures that abiotic factors are the

dominant factors controlling morphology. All the algal groups can show similar morphologies.

The second contribution is the recognition of biogenetic processes in the formation of mottled dolomitic limestones. Although Jones *et al.* (1979) addressed the problem of the genesis of rubbly limestones on Somerset Island and Narbonne (1981, 1984) the systematic ichnology of these rocks; a synthesis incorporating the role of bioturbation in the genesis of these rocks has been lacking. The Upper Silurian mottled dolomitic limestones on eastern Prince of Wales Island show a similar biogenetic and diagenetic history as the Cretaceous and Jurassic chalks of Europe. With minor revisions, this model is directly applicable to these mottled dolomitic limestones and demonstrates a genetic link between the three types of mottled dolomitic limestones. The application of the model for Cretaceous and Jurassic chalks also demonstrates the law of uniformitarianism.

The third contribution is the delineation of the role the Boothia Horst played as a sediment source and as a control on facies relationships in the Ludlovian. The Boothia Horst is acknowledged as the single largest influence on the geological history of the area. Two aspects of its influence became apparent during this study. One is the abundance of detrital dolomites in the Cape Storm and Somerset Island formations. Although the question of provenance remains unresolved, much of the dolomites are

possibly terrigenous and derived from strata onlapping the Boothia Horst. During times of terrigenous influx into the M'Clintock Basin, the Boothia Horst was a low-lying emergent feature.

The second aspect of the Boothia Horst was evident during subsidence when it was apparent that the M'Clintock Basin was segmented into five quasi-independent sub-basins along its eastern margin. Differential subsidence of fault blocks along contemporaneous basement faults is the mechanism attributed to the formation of the sub-basins. Variations in the regional facies of the Upper Silurian strata on eastern Prince of Wales Island are associated with these sub-basin.

Independently, the lines of evidence indicate the M'Clintock Basin was a shallow-subtidal basin with slow rates of sedimentation and extensive tidal flats. Collectively, the evidence shows the M'Clintock Basin to be an excellent example of an ancient tidal flat to shallow subtidal sequence. The transgressive-regressive cycle is relatively symmetrical with the regressive phase approximately an inversed sequence of the transgressive phase. Interpretation of the Upper Silurian strata on eastern Prince of Wales Island was enhanced by examining the full transgressive-regressive couplet rather than a single formation within this cycle.

As is the situation with most studies, at least as many questions are posed as resolved. This study is no exception

and the following studies may prove useful in the understanding of the geological history of this area;

1. There are basic problems in the stratigraphic nomenclature of the Upper Silurian strata proximal to the Boothia Horst. Detailed petrographic and stratigraphic work between and within these basins may resolve some of these problems.
2. The provenance of the detrital dolomites. Geochemical and cathodalumiscent microscopy may be useful to prove both the provenance and diagenesis of these dolostones of the Cape Storm Formation and underlying strata. A regional study of the Cape Storm Formation may delineate the geographic distribution of detrital dolomites relative to the Boothia Horst.
3. Re-evaluation of the mottled dolomitic limestones of the Douro Formation of the Arctic Lowlands to check whether the biogenetic origin is endemic to eastern Prince of Wales Island or cosmopolitan.

IX. PHOTOGRAPHIC PLATES

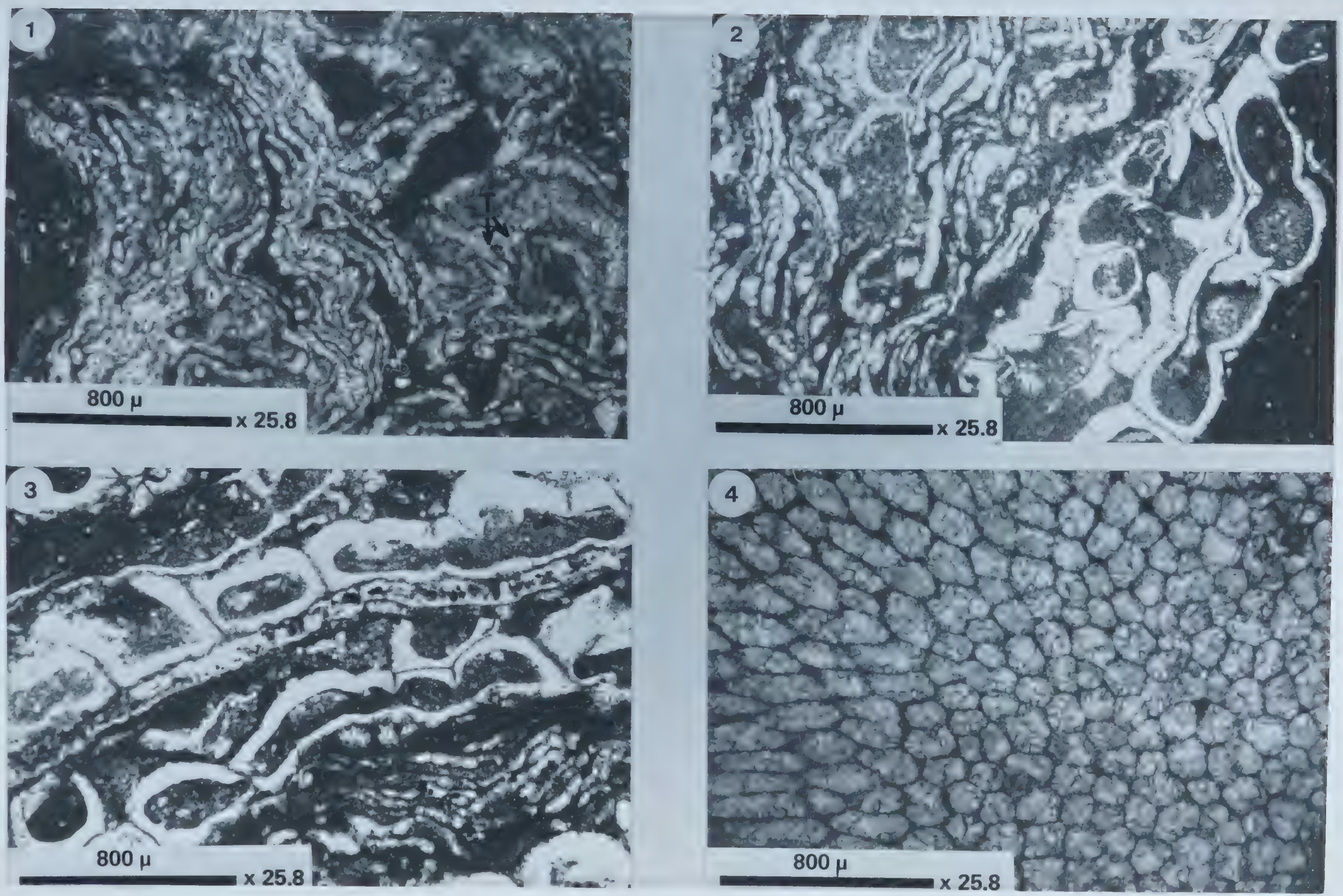
PLATE 1

Figure 1. Photomicrograph of *Sphaerocodium* sp. and *Girvanella* sp.. The smaller filaments are *Girvanella* sp. and the larger beaded filaments are *Sphaerocodium* sp. T=tangential section of *Sphaerocodium* sp. showing the characteristic fanlike branching.
Unit M-56, lower part of the Somerset Island Formation.

Figure 2. Photomicrograph of intergrowths of *Sphaerocodium* sp. and *Girvanella* sp.. The larger, bilobate filaments on the right side of the figure are *Wetheredella sp.*
Unit M-56, lower part of the Somerset Island Formation.

Figure 3. Photomicrograph of *Wetheredella* sp. filaments. The smaller filaments are *Sphaerocodium* sp.
Unit M-56, lower part of the Somerset Island Formation.

Figure 4. Photomicrograph of *Solenopora* sp. in transverse section. The cells are hexagonal in outline and radiate outwards.
Unit Q-55, lower part of the Somerset Island Formation.



1
800 µ
x 25.8
2
800 µ
x 25.8
3
800 µ
x 25.8
4
800 µ
x 25.8

PLATE 2

Figure 1. Polished slab of a spherical oncolith (Type A). A diagenetic halo occurs around the oncolith.
Unit M-73, Somerset Island Formation.

Figure 2. Negative print of the oncolith in Figure 1.
Unit M-73, Somerset Island Formation.

Figure 3. Photomicrograph of the outer margins of the oncolith in Figure 1. Note the random orientation of the *Girvanella* sp. filaments.
Unit M-73, Somerset Island Formation.

Figure 4. Photomicrograph of the oncolith in Figure 1. The larger filaments in the right side of the figure are *Wetheredella sp.*
Unit M-73, Somerset Island Formation.

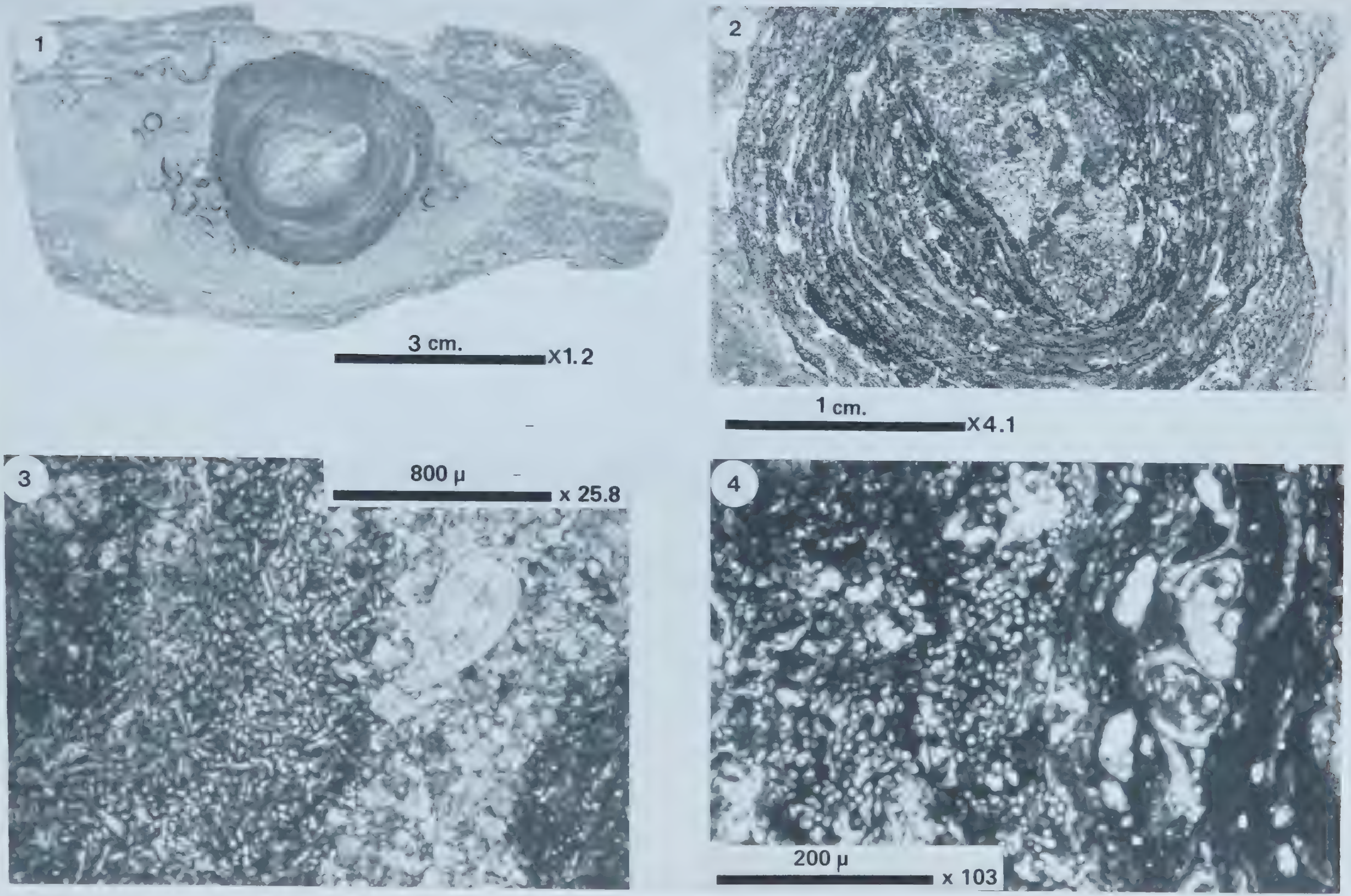
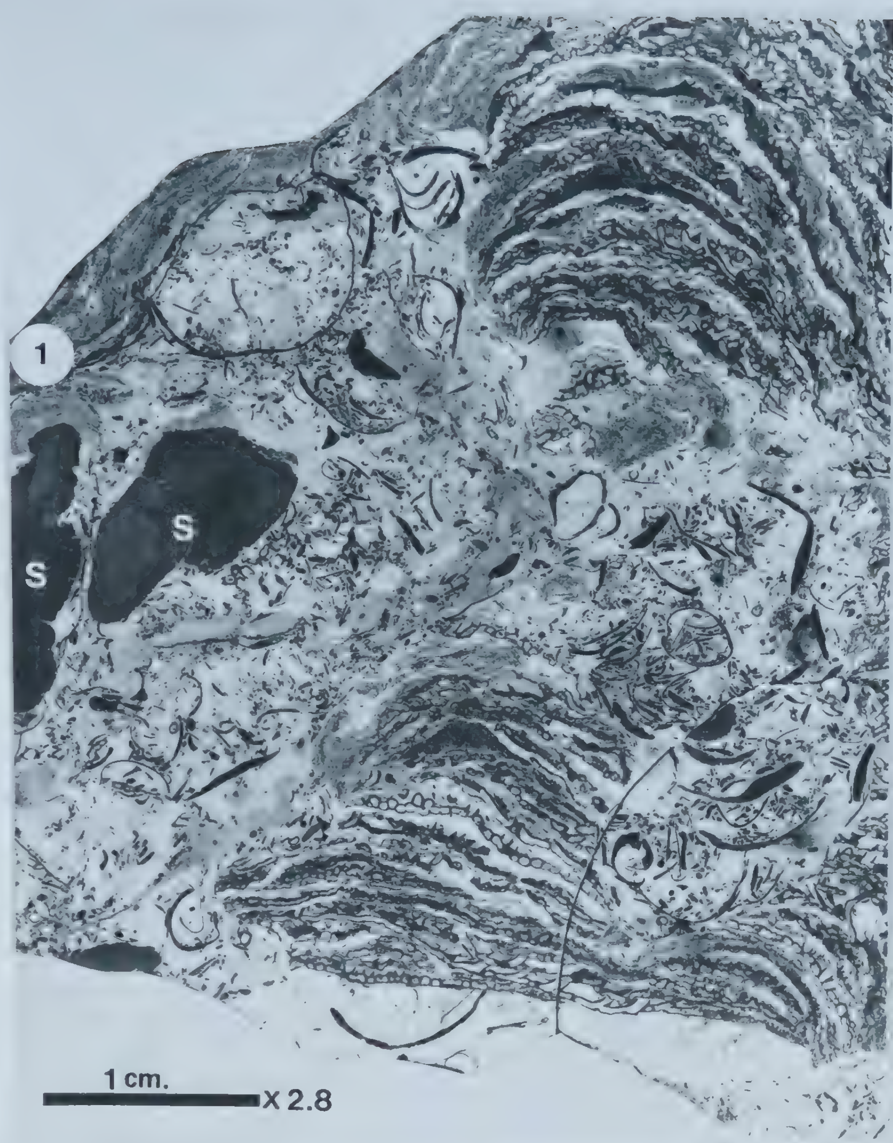

1
3 cm.
X1.2
2
1 cm.
X4.1
3
800 μ
x 25.8
4
200 μ
x 103

PLATE 3

Figures 1 and 2. Negative print of the internal structure of the algal columns. The larger filaments are *Wetheredella* sp. and the smaller filaments are *Girvanella* sp. and *Sphaerocodium* sp.. S=*Solenopora filiformis* nodules (=rhodoliths). Note the algal encrustations on the rhodoliths and the presence of a *Girvanella* mat in the upper left corner of Figure 1.
Unit M-56, lower part of the Somerset Island Formation.


1
S
S
1 cm.
X 2.8

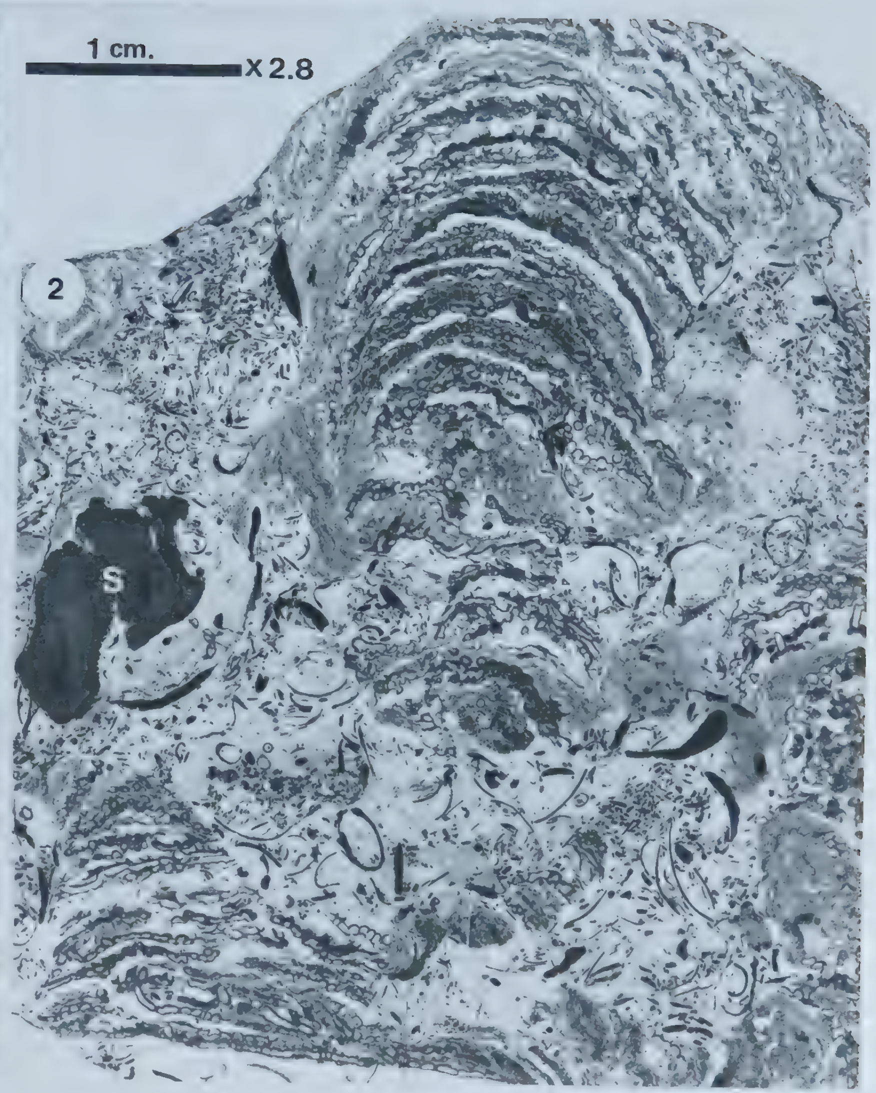
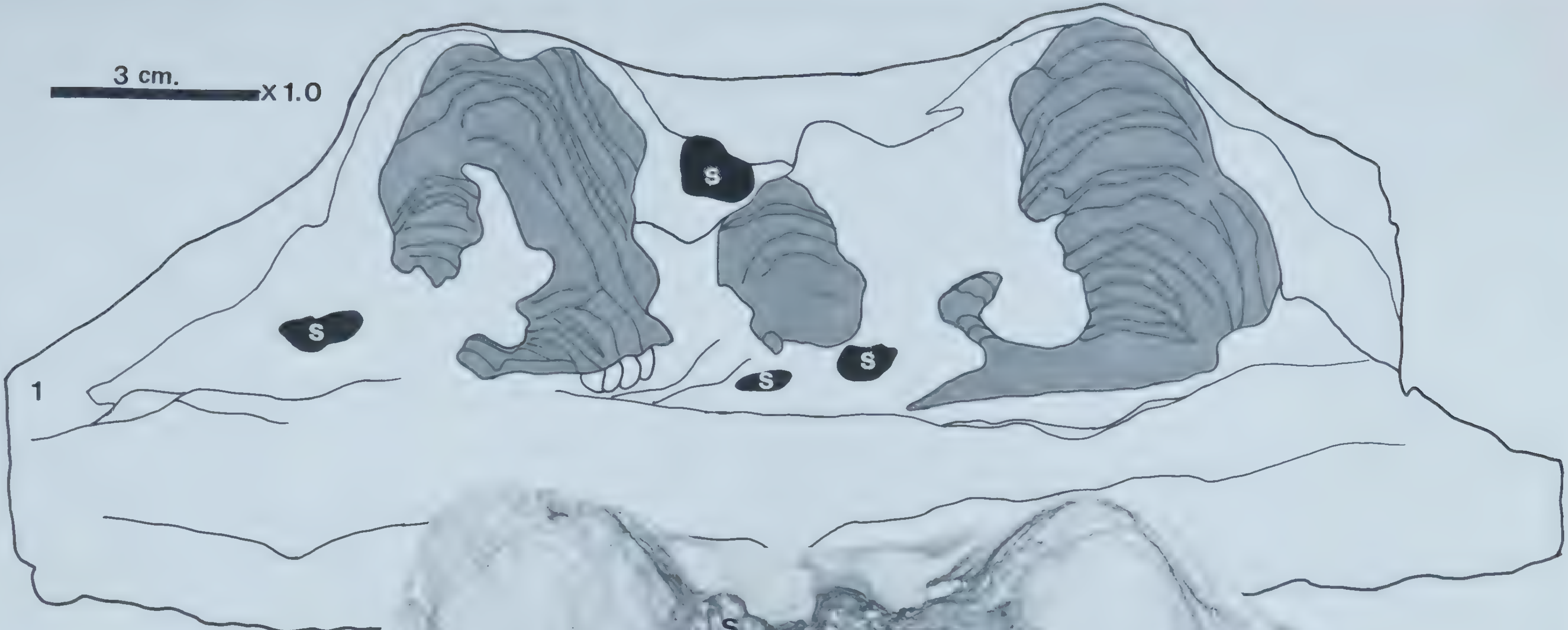

2
S
1 cm.
X 2.8

PLATE 4

Figure 1. Line drawing of the algal columns. The light shaded areas are the algal columns and the black shaded areas (S) are *Solenopora*.
Unit M-56, lower part of the Somerset Island Formation.

Figure 2. Polished slab of the algal columns in Figure 1. S=*Solenopora*.
Unit M-56, lower part of the Somerset Island Formation.

1


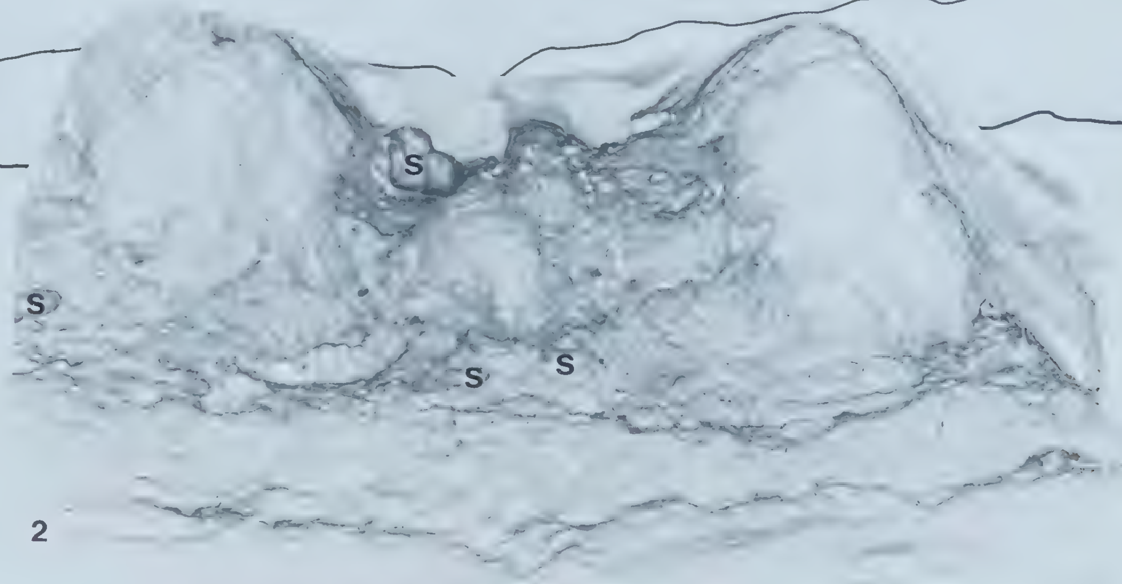
2


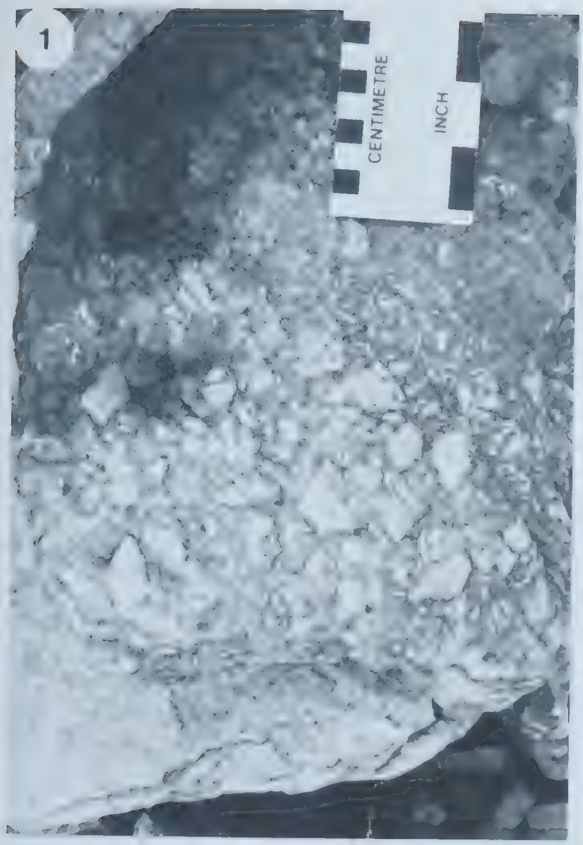
PLATE 5

Figure 1. Field photograph of a dolosiltite intraclast breccia.
Unit L-20, Cape Storm Formation.

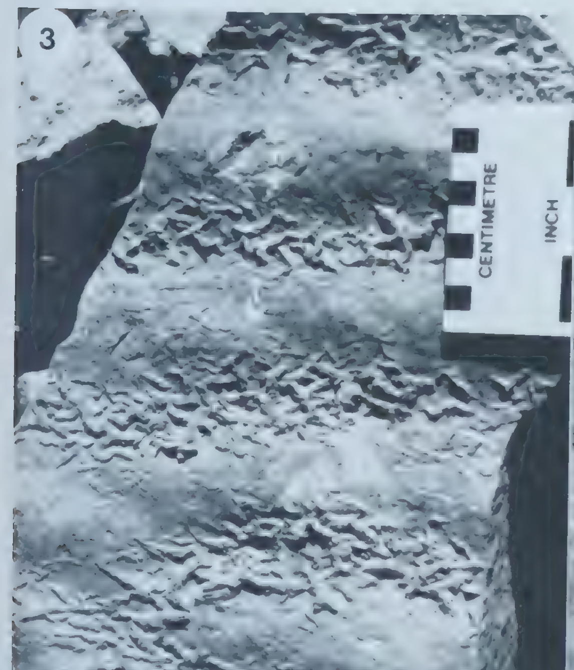
Figure 2. Field photograph of a micrite intraclast breccia.
Unit M-48, lower part of the Somerset Island Formation.

Figure 3. Field photograph of ripple marks and bioturbation in the troughs.
Umit M-69, Somerset Island Formation.

Figure 4. Recent analogue of Figure 3. Intertidal mudflat, Cresent Beach, British Columbia. Lower tidal flat zone of Kellerhals and Murray (1976). Cresent Beach is part of the Boundary Bay tidal flat study of the above authors.


1
CENTIMETRE
INCH


2
10 cm.


3
CENTIMETRE
INCH


4

PLATE 6

Figure 1. Field photograph of desiccation polygons in a dolomitic micrite.
Unit M-46, lower part of the Somerset Island Formation.

Figure 2. Field photograph of a cast of synaeresis cracks in a rippled dolomitic micrite.
Unit M-44, lower part of the Somerset Island Formation.

Figure 3. Field photograph of desiccation polygons in a quartzose micrite.
Unit N-65, lower part of the Somerset Island Formation.

Figure 4. Field photograph of intraclasts showing some imbrication.
Unit A-74, lower part of the Somerset Island Formation.

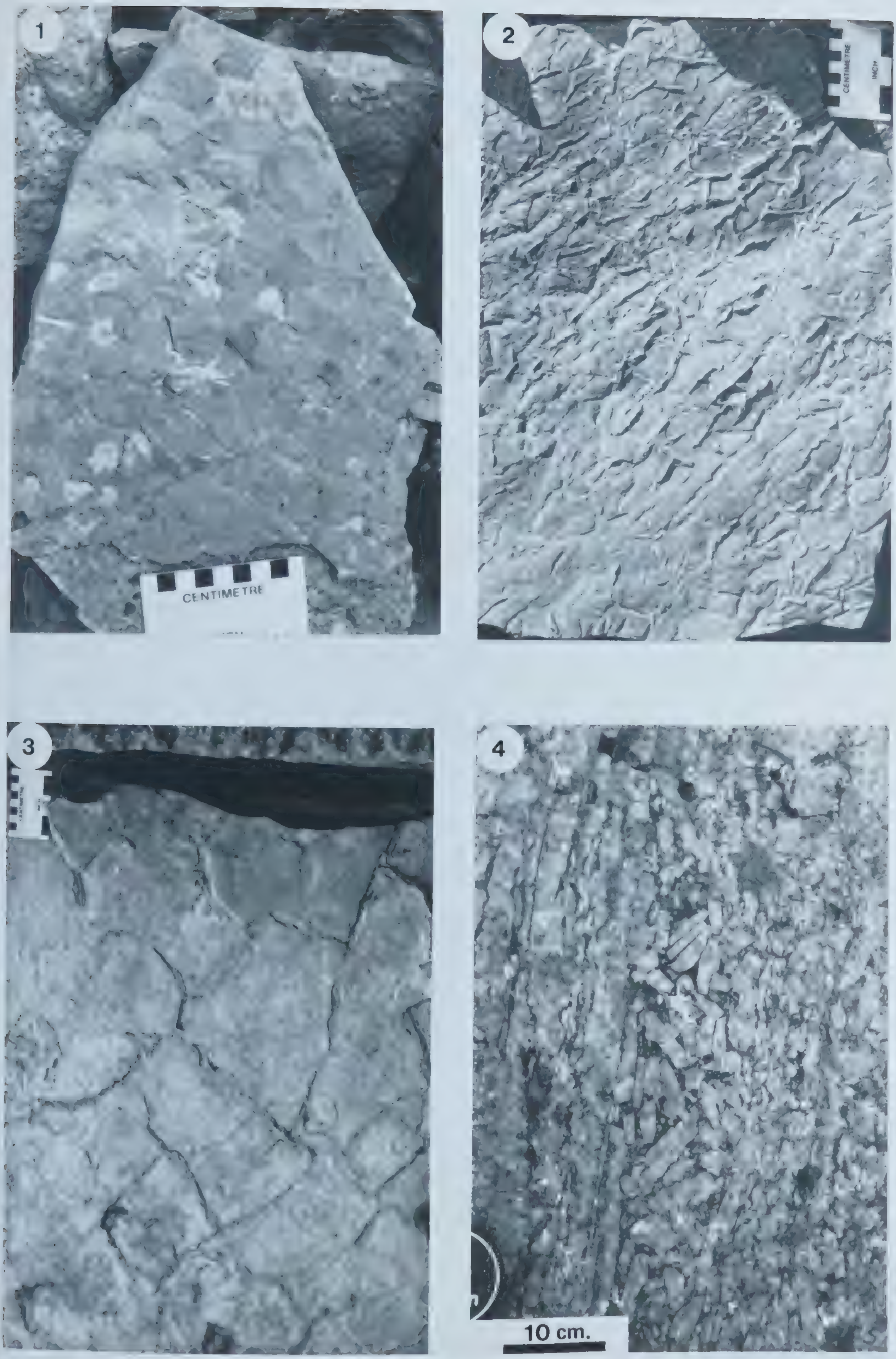

1
CENTIMETRE
2
3
4
10 cm.

PLATE 7

Figure 1. Field photograph of hemispherical stromatolites along strike.
Unit M-45, lower part of the Somerset Island Formation.

Figure 2. Field photograph of hemispherical stromatolites weathering out of a dolosiltite.
Unit D-46, lower part of the Somerset Island Formation.

Figure 3. Field photograph of oncoliths on a bedding plane.
Unit R-61, Somerset Island Formation.

Figure 4. Field photograph of stacked hemispherical stromatolites.
Unit B-12, member B, Cape Storm Formation.

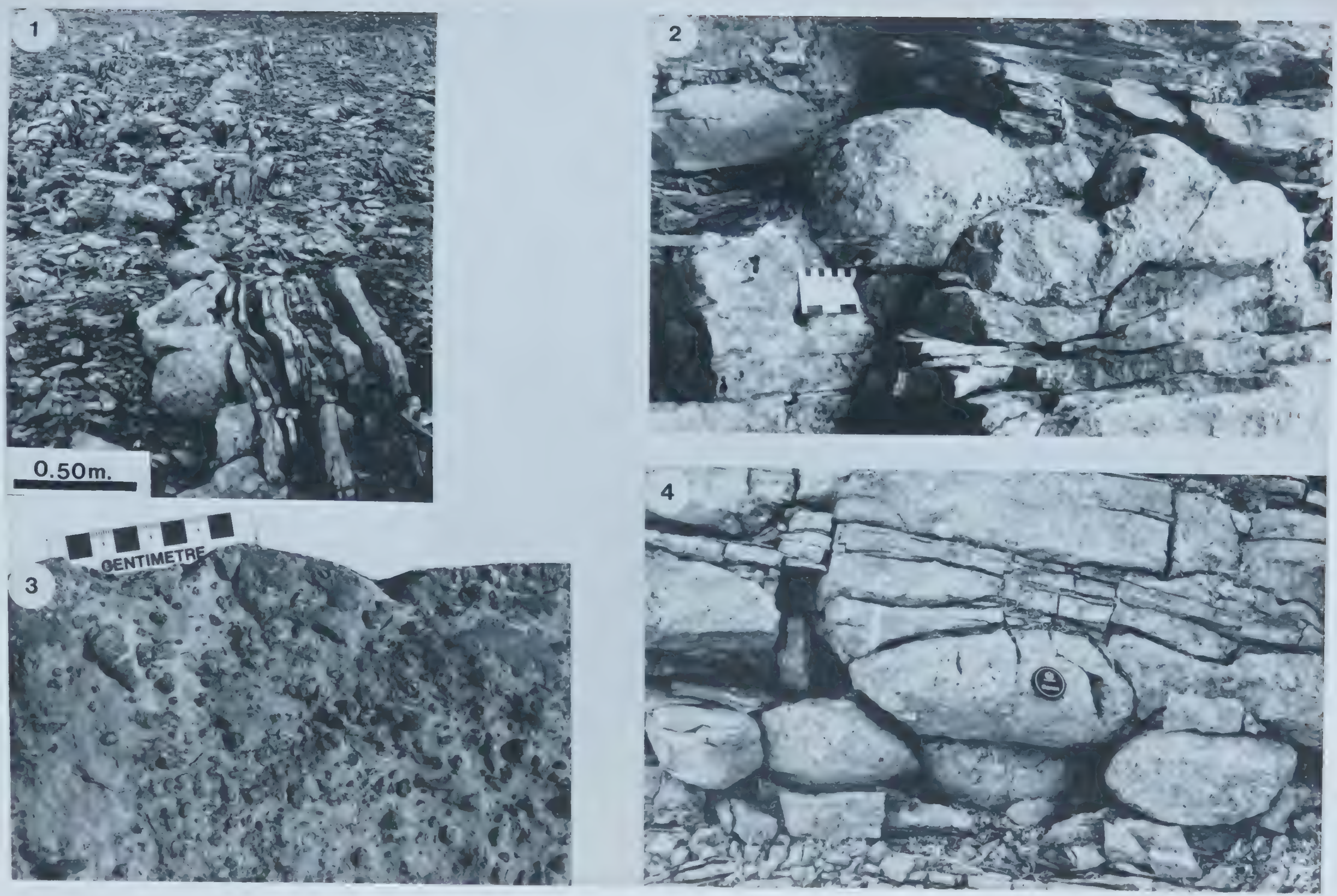

1
0.50m.
2
3
CENTIMETRE
4

PLATE 8

Figure 1. Photomicrograph of *Vermiporella sp.*, oblique section.
Unit K-23, upper Douro Formation.

Figure 2. Photomicrograph of *Vermiporella sp.*, oblique and tangential sections.
UNit K-23, upper Douro Formation.

Figure 3. Photomicrograph of *Girvanella problematica*.
Unit K-23, upper Douro Formation.

Figure 4. Photomicrograph of *Girvanella problematica*. The circular cells in the right side of the photograph may be in the reproductive stage.
Unit K-23, upper Douro Formation.

Figure 5. Photomicrograph of *Girvanella problematica*.
Unit K-23, upper Douro Formation.

Figure 6. Photomicrograph of *Girvanella problematica*. The row of circular cells in the lower right corner of the photomicrograph may be in the reproductive stage.
Unit K-23, upper Douro Formation.

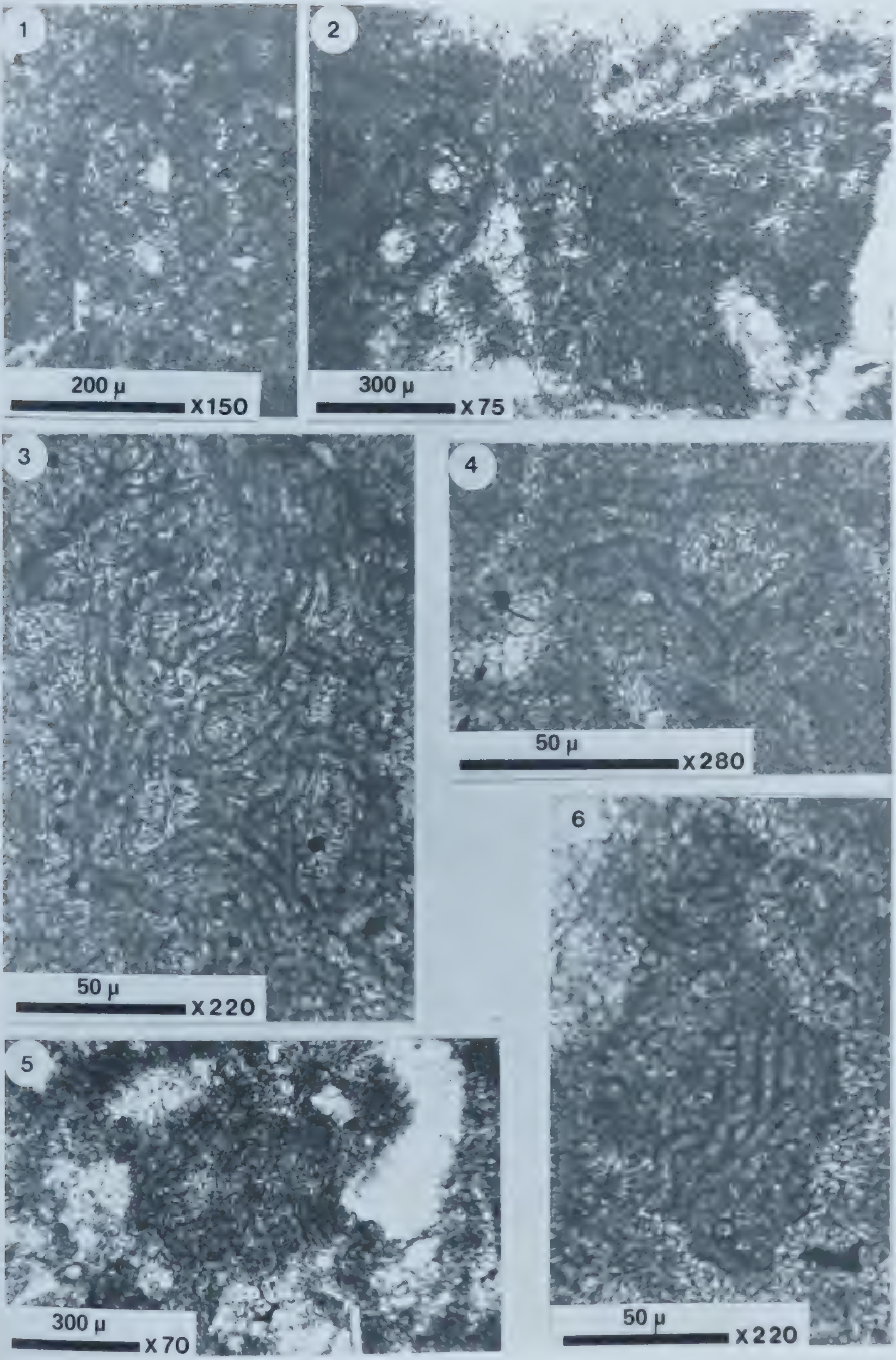

1
200 μ
X150
2
300 μ
X75
3
50 μ
X220
4
50 μ
X280
5
300 μ
X70
6
50 μ
X220

PLATE 9

Figure 1. Field photograph of Type B oncolith. Note the lack of laminae in the algal overgrowths.
Unit Q-44, lower part of the Somerset Island Formation.

Figure 2. Photomicrograph of *Girvanella* sp. filaments in the algal overgrowth in Figure 1. Note the random orientation of the filaments.
Unit Q-44, lower part of the Somerset Island Formation.

Figure 3. Photomicrograph of *Girvanella* sp. filaments in the algal overgrowths in Figure 1.
Unit Q-44, lower part of the Somerset Island Formation.

Figure 4. Negative print of Type D oncoliths. The nucleus is composite, composed of a lithoclast of different composition than the surrounding strata and of a *Solenopora* nodule.
Unit Q-44, lower part of the Somerset Island Formation.

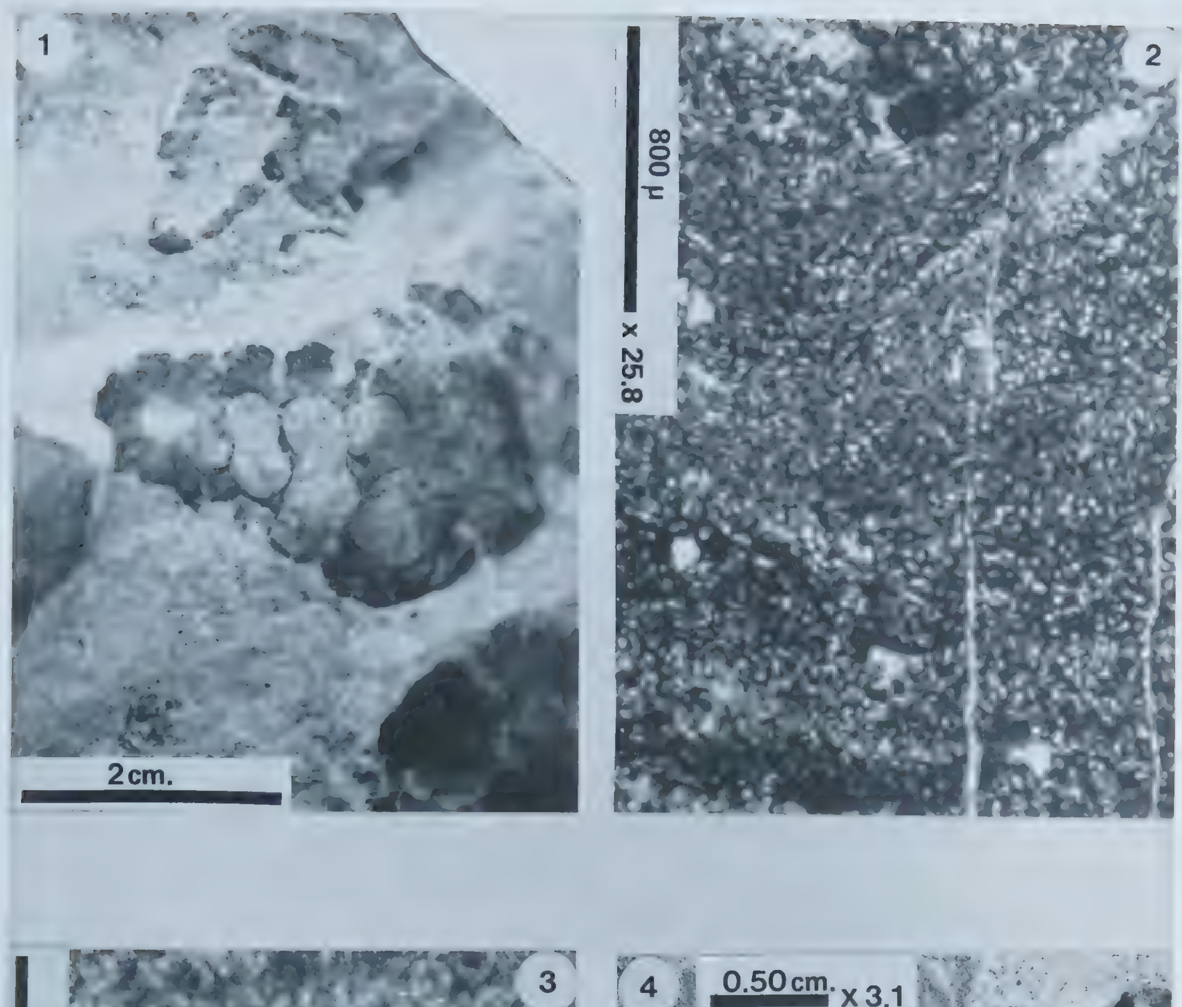

1
2 cm.
2
800 µ
x 25.8
3
300 µ
x 65
4
0.50 cm.
x 3.1

PLATE 10

Figure 1. Negative print of a Type D lobate oncolith. Unit Q-38, lower part of the Somerset Island Formation.

Figure 2. Negative print of a Type D lobate oncolith. The core is spherical with a later lobate growth. Note the planar erosion surface truncating the oncolith. Unit R-59, lower part of the Somerset Island Formation.

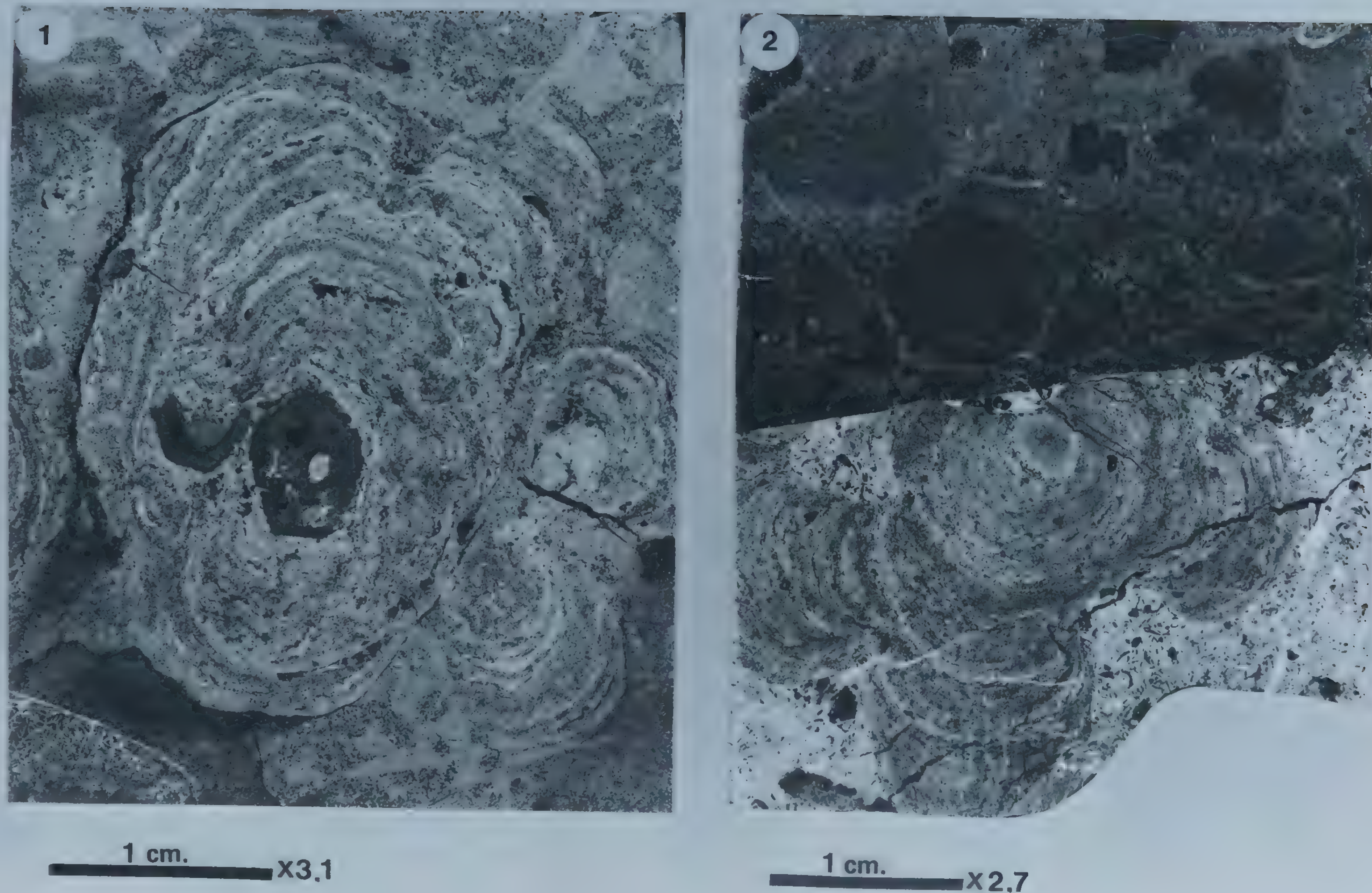

1
2
1 cm.
X3.1
1 cm.
X 2.7

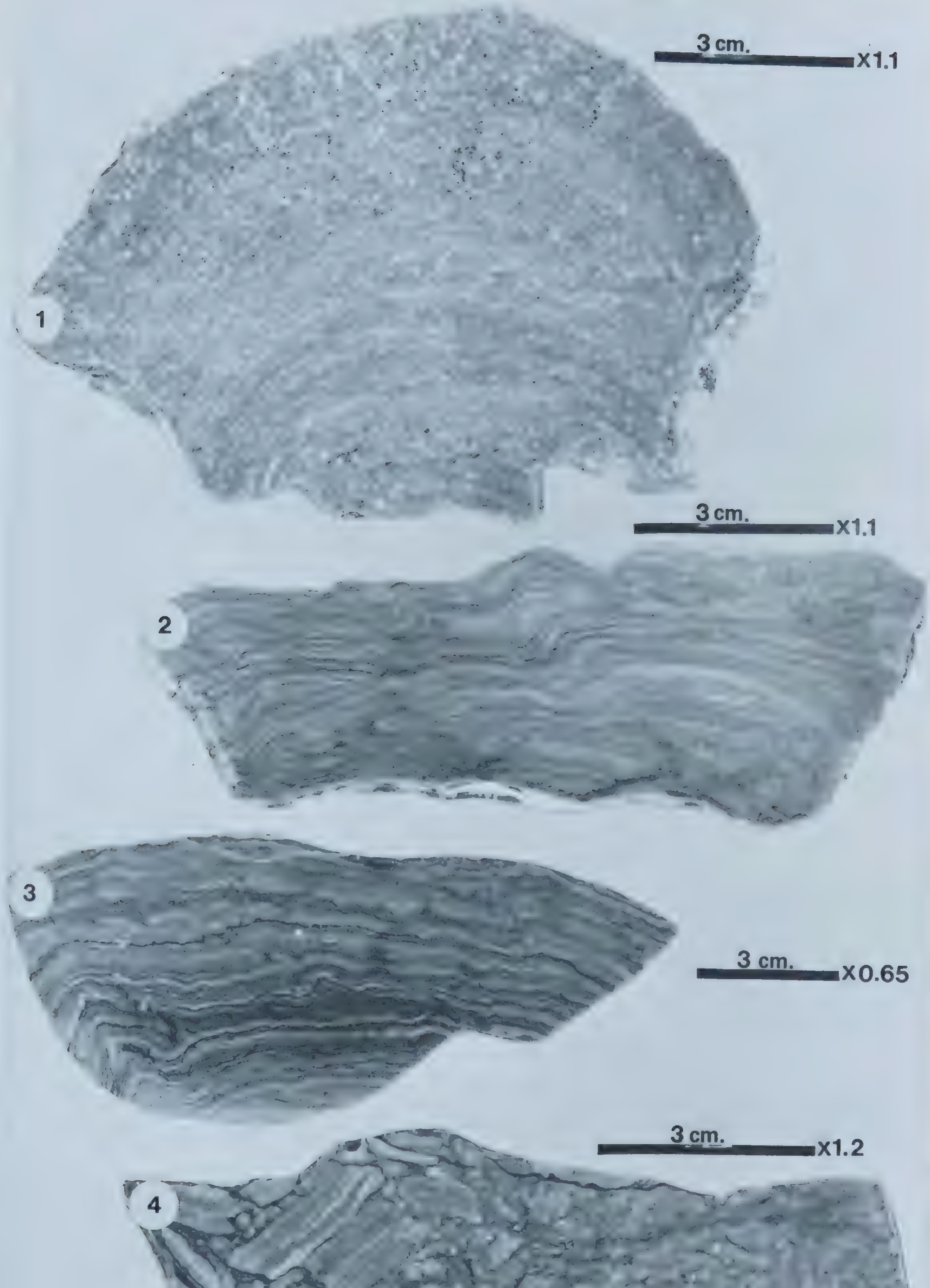
PLATE 11

Figure 1. Polished slab of a cross-sectional view of a hemispherical stromatolite. The dark spots are ooids entrapped in fractures.
Unit C-22, member B, Cape Storm Formation.

Figure 2. Polished slab of an undulatory stromatolite.
Unit C-22, member B, Cape Storm Formation.

Figure 3. Polished slab of an undulatory stromatolite.
Unit G-40, member B, Cape Storm Formation.

Figure 4. Polished slab of a brecciated algal laminite. Sample occurs in the interarea between hemispherical stromatolites.
Unit G-37, member B, Cape Storm Formation.



1
3 cm.
X1.1
2
3 cm.
X1.1
3
3 cm.
X0.65
4
3 cm.
X1.2

PLATE 12

Figure 1. Polished slab of digitate stromatolites.
Unit L-29, member B, Cape Storm Formation.

Figure 2. Line drawing of the digitate stromatolites in Figure 1.
Unit L-29, member B, Cape Storm Formation.

Figure 3. Negative print of a digit from the digitate stromatolite in Figure 1. showing the characteristic laminae of stromatolites.
Unit L-29, member B, Cape Storm Formation.

Figure 4. Photomicrograph of digit in Figure 3. Note the absence of microstructure and obscurity of laminae under high magnification.
Unit L-29, member B, Cape Storm Formation.

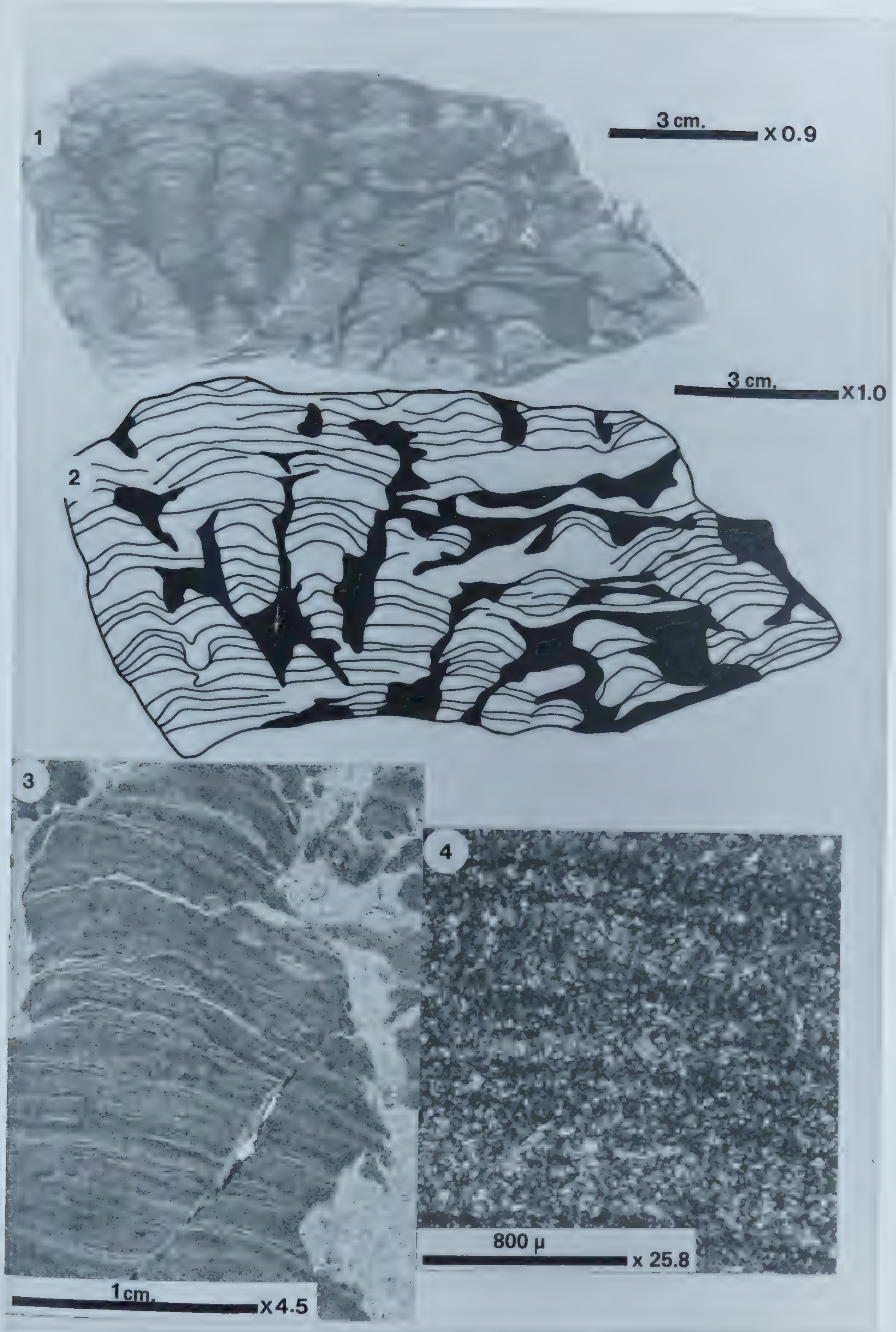
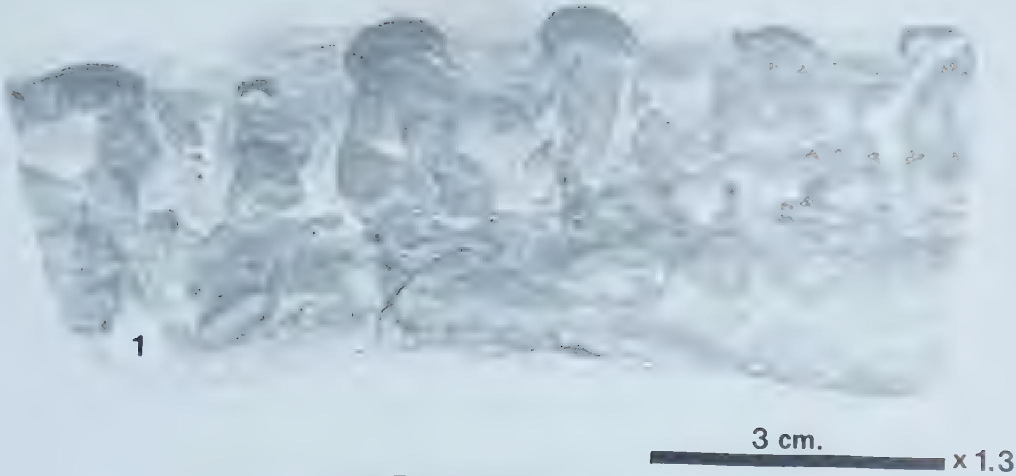

1
3 cm.
x 0.9
3 cm.
x1.0
2
3
4
800 μ
x 25.8
1cm.
x4.5

PLATE 13

Figure 1. Polished slab of columnar thrombolites.
Unit A-66, upper Douro Formation.

Figure 2. Line drawing of thrombolites in Figure 1. showing the four algal growth phases. The top of phase 4 and the thin algal trace joining phase 4 represents original depositional topography.
Unit A-66, upper Douro Formation.

Figure 3. Negative print of the thrombolites in Figure 1. showing the clotted fabric and laminated dolosiltites.
Unit A-66, upper Douro Formation.


1
3 cm.
x 1.3


4
3
2
1.
2
3 cm.
x 1

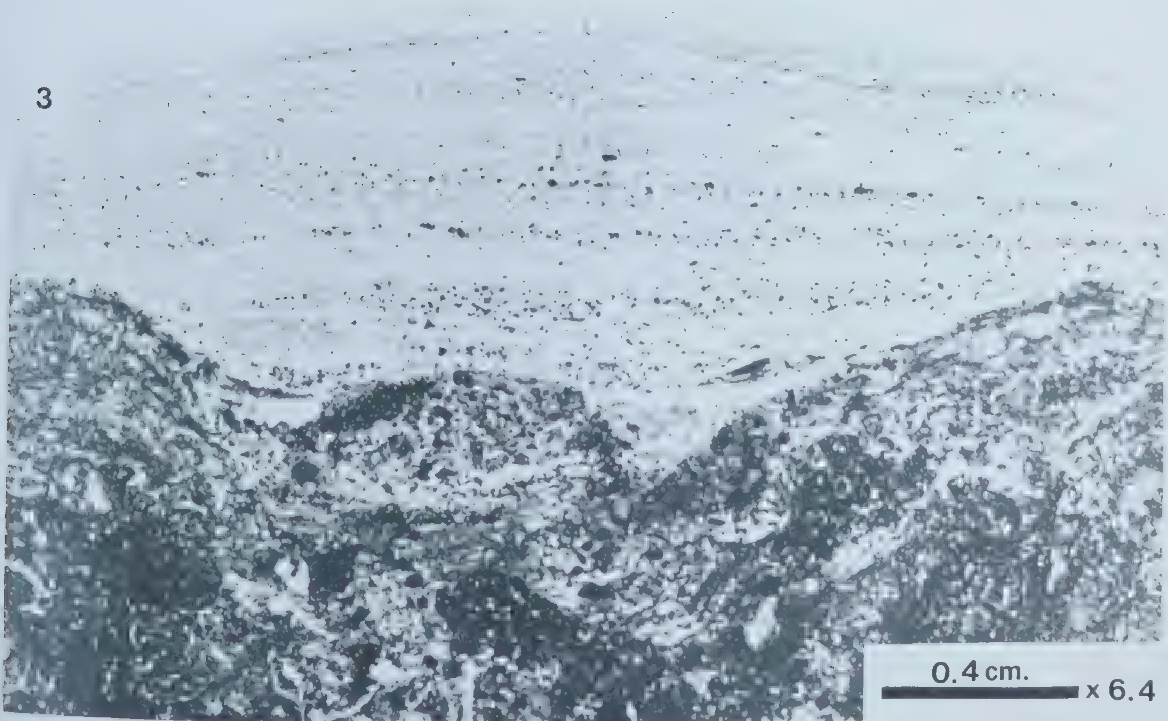

3
0.4 cm.
x 6.4

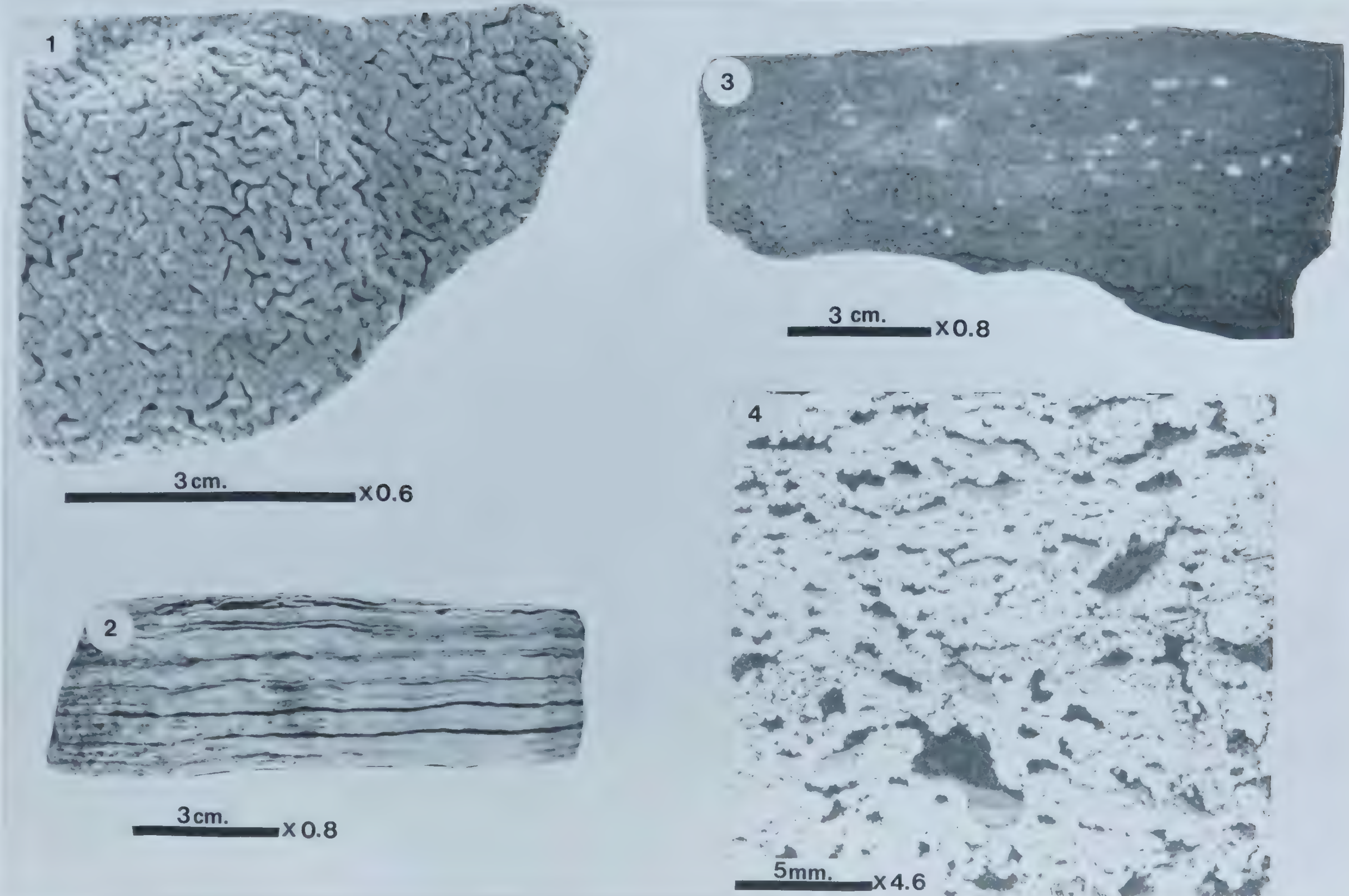
PLATE 14

Figure 1. Hand sample of a colloform algal laminite (mat) in a dolomitic calcisiltite, bedding plane view. Error in magnification, should be x1.6 not x0.6.
Unit Q-67, Somerset Island Formation.

Figure 2. Hand sample of a planar algal laminite, cross-sectional view.
Unit N-94, Somerset Island Formation.

Figure 3. Polished slab of fenestrae in a micrite. Fenestrae define crude laminations.
Unit N-35, Douro Formation.

Figure 4. Negative print of fenestrae in Figure 3. Note the partial filling by calcisiltites with the remainder of the void infilled by sparry calcite.
Unit N-35, Douro Formation.



1
3 cm.
X 0.6
2
3 cm.
X 0.8
3
3 cm.
X 0.8
4
5 mm.
X 4.6

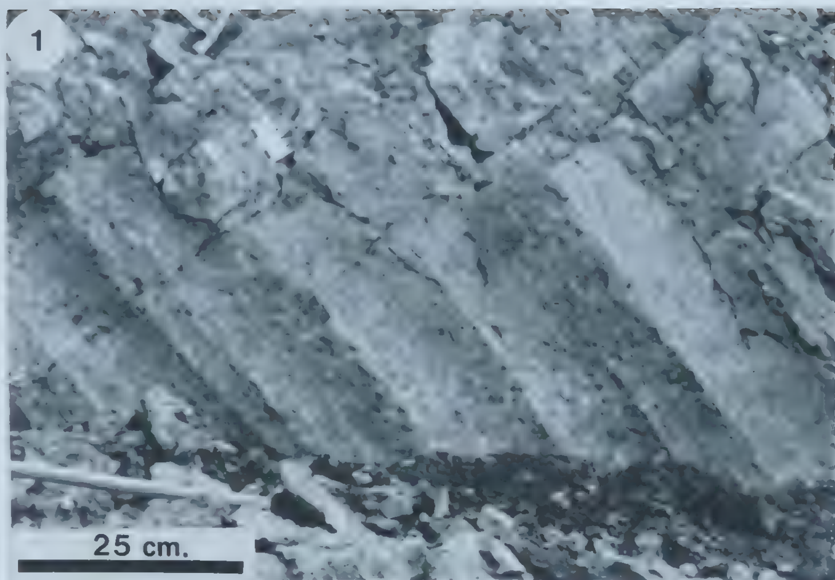
PLATE 15

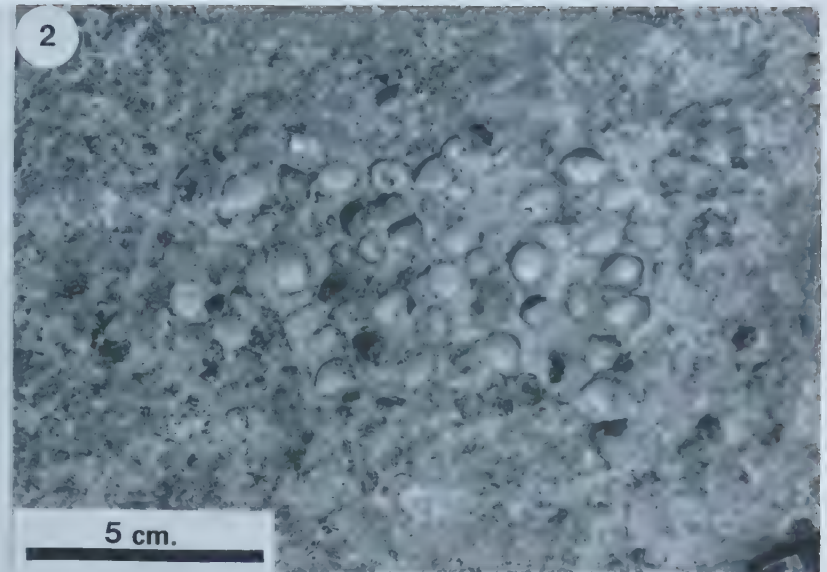
Figure 1. Field photograph of ripple marks at the top of an algal laminite unit.
Unit N-35, Douro Formation.

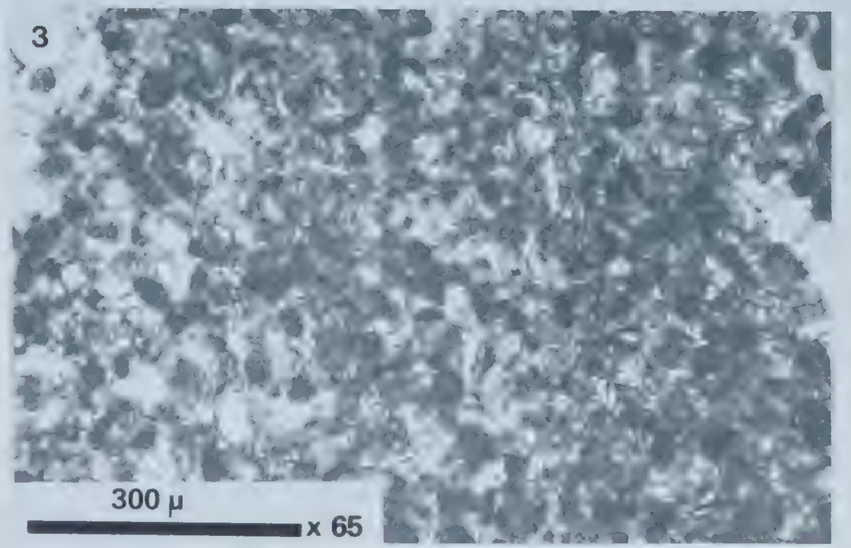
Figure 2. Field photograph of *Atrypoidea erebus* on a bedding plane in a Type III mottled dolomitic biomicrite.
Unit Q-55, lower part of the Somerset Island Formation.

Figure 3. Photomicrograph of a pelmicrite in a mottled dolomitic biopelmicrite.
Unit, A-65, upper Douro Formation.

Figure 4. Field photograph of a Type III mottled dolomitic biomicrite (unit Q-68) and a Type II mottled dolomitic biomicrite (unit Q-69).
Units Q-68 and 69, Somerset Island Formation.


1
25 cm.


2
5 cm.


3
300 µ
x 65

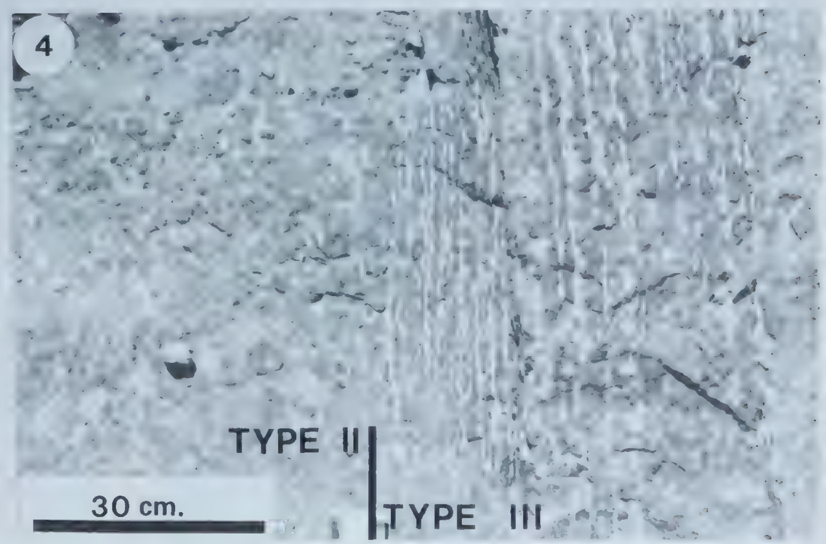
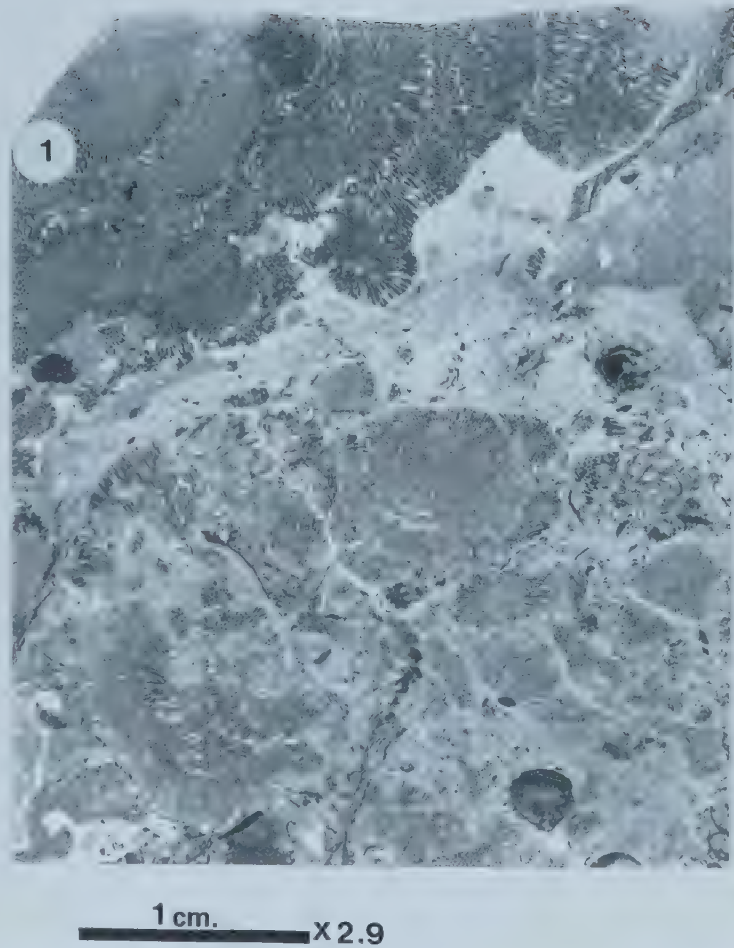

4
TYPE II
TYPE III
30 cm.

PLATE 16

Figure 1. Negative print of *Solenopora sp.* in an algal biolithite.
Unit Q-55, lower part of the Somerset Island Formation.

Figure 2. Negative print of *Solenopora sp.* in an algal biolithite.
Unit R-62, lower part of the Somerset Island Formation.


1
1 cm.
X 2.9

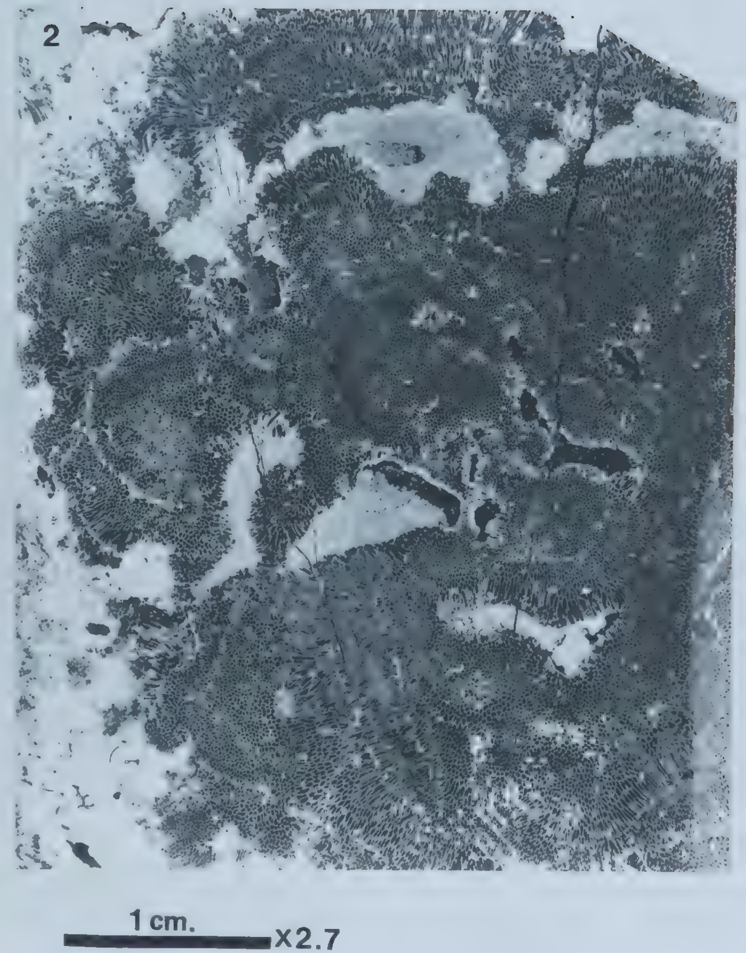

2
1 cm.
X2.7

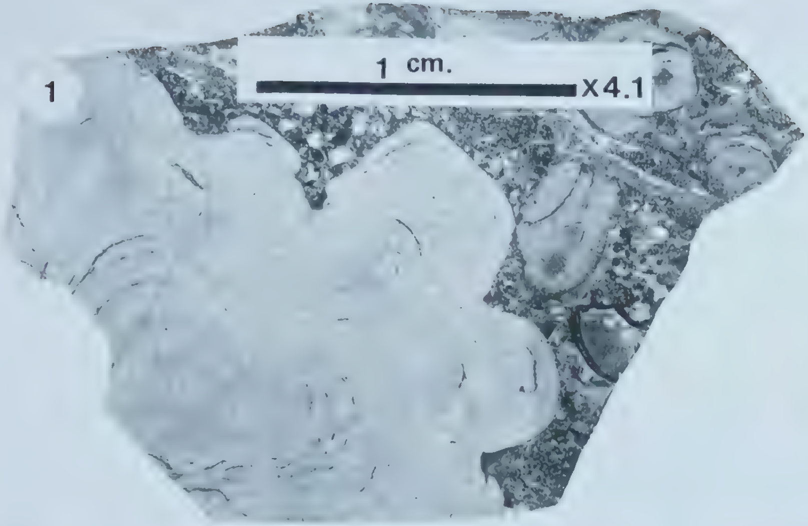
PLATE 17

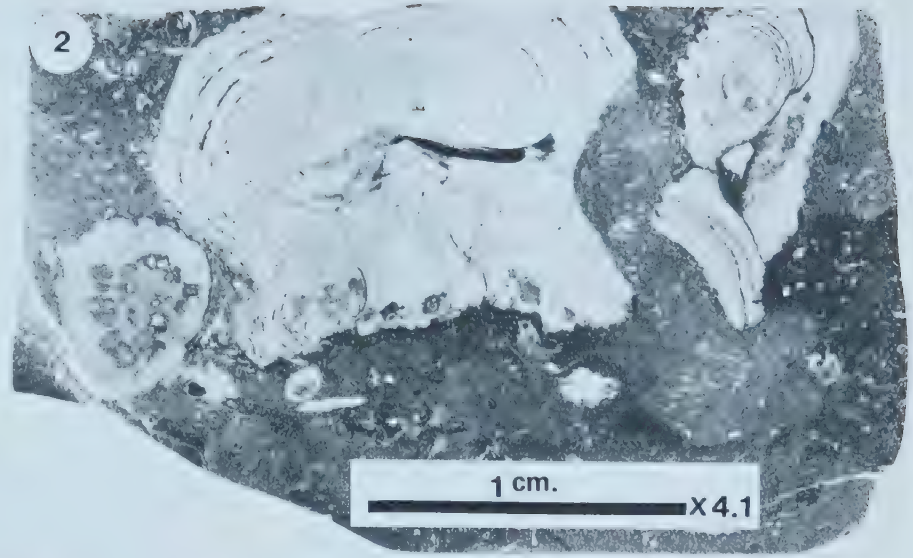
Figure 1. Negative print of a mammalated rhodolith comprised of *Solenopora filiformis* in a dolomitic micrite.
Unit M-55, lower part of the Somerset Island Formation

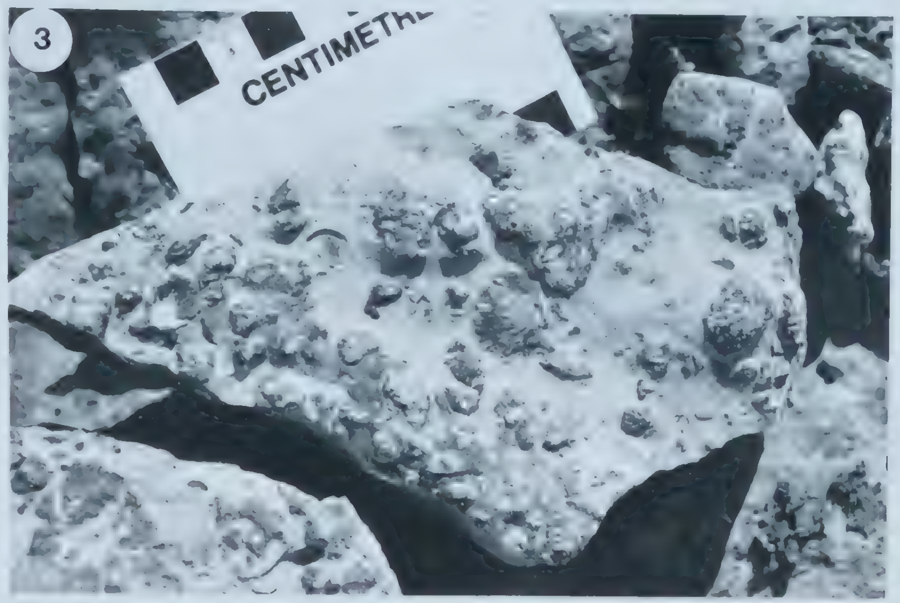
Figure 2. Negative print of a rhodolith comprised of *Solenopora filiformis*. The base of the rhodolith appears "torn" suggesting the rhodolith was possibly once attached to the subtrate.
Unit G-104, Somerset Island Formation.

Figure 3. Field photograph of rhodoliths on a bedding plane. Weathering accentuates the radial nature of the cell threads allowing for identification of *Solenopora* in the field.
Unit R-59, lower part of the Somerset Island Formation.

Figure 4. Photomicrograph of a vertical section of *Solenopora filiformis* showing the radiating cell threads. Note the dominance of the cell threads relative to the cross partitions.
Unit M-55, lower part of the Somerset Island Formation.


1
1 cm.
X4.1


2
1 cm.
X4.1


3
CENTIMETR

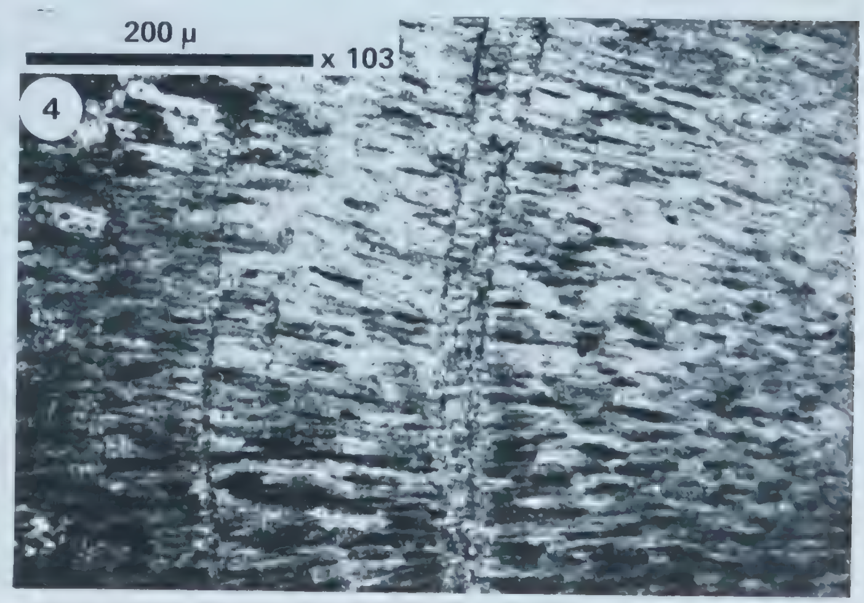

4
200 µ
x 103

PLATE 18

Figure 1. Hand sample of *Solenopora* nodules on a bedding plane. b=*Atrypoidea netserki*. The other spheroids are rhodoliths.
Unit G-109, Somerset Island Formation.

Figure 2. Polished slab showing grain diminution of *Solenopora* sp. to micrite. a= recognizable *Solenopora* sp. fragments grading into a featureless micrite.
Unit G-104, Somerset Island Formation.

Figure 3. Photomicrograph of a sparry calcite cemented, skeletal calcarenite. S=*Solenopora* nodules.
Unit B-59, Douro Formation.

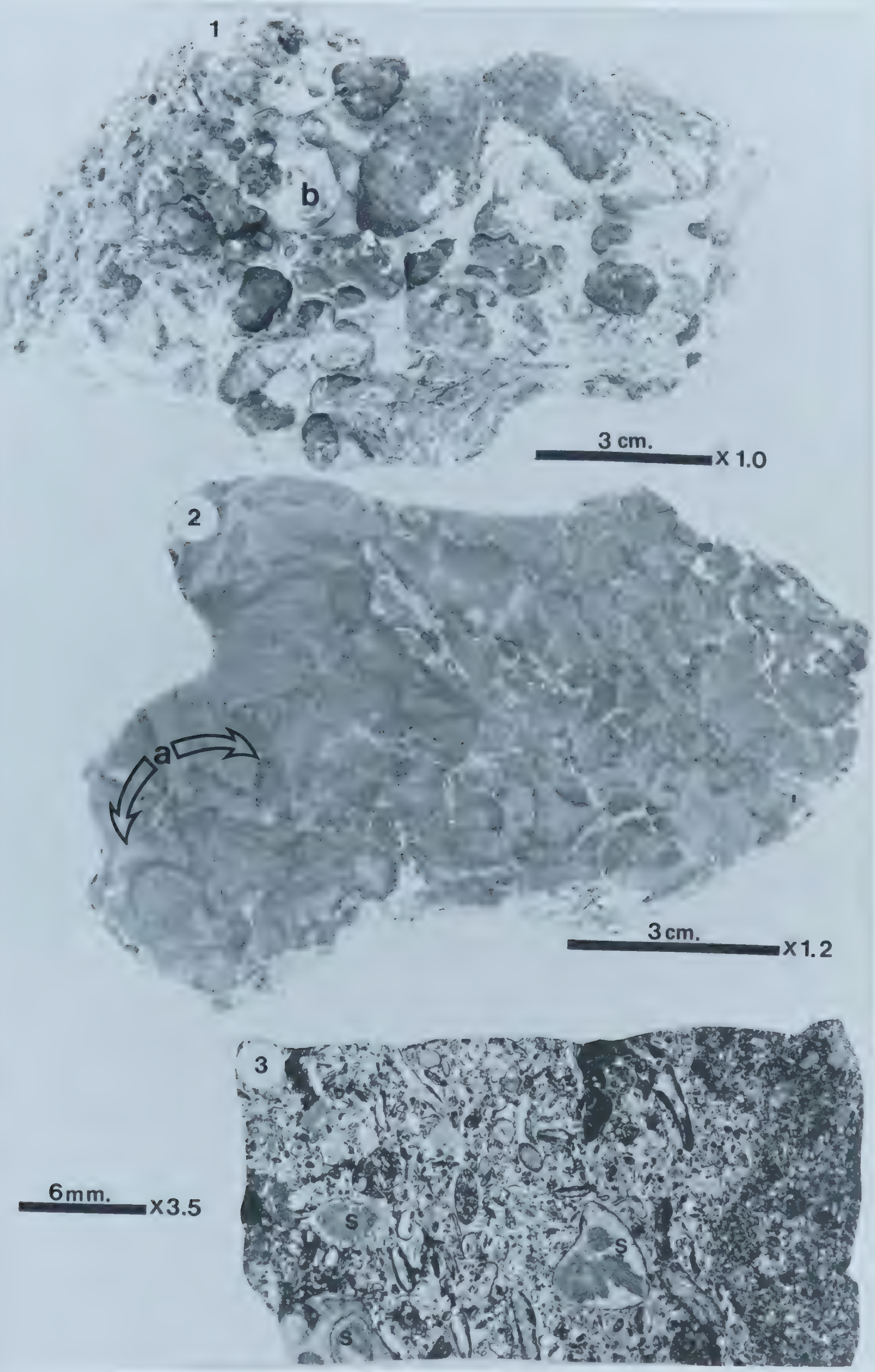

1
b
3 cm. x 1.0
2
a
3 cm. x1.2
3
6mm. x3.5
S
S
S

PLATE 19

Figure 1. Field photograph of burrowing in a rubbly dolomitic limestone. Exposure is in covex hyporelief. The larger burrows (T) are possibly *Thalassinoides* and the smaller burrows (C) are *Chondrites*. This ichnoassemblage is possibly of the preomission suite.
Unit G-75-b, Douro Formation.

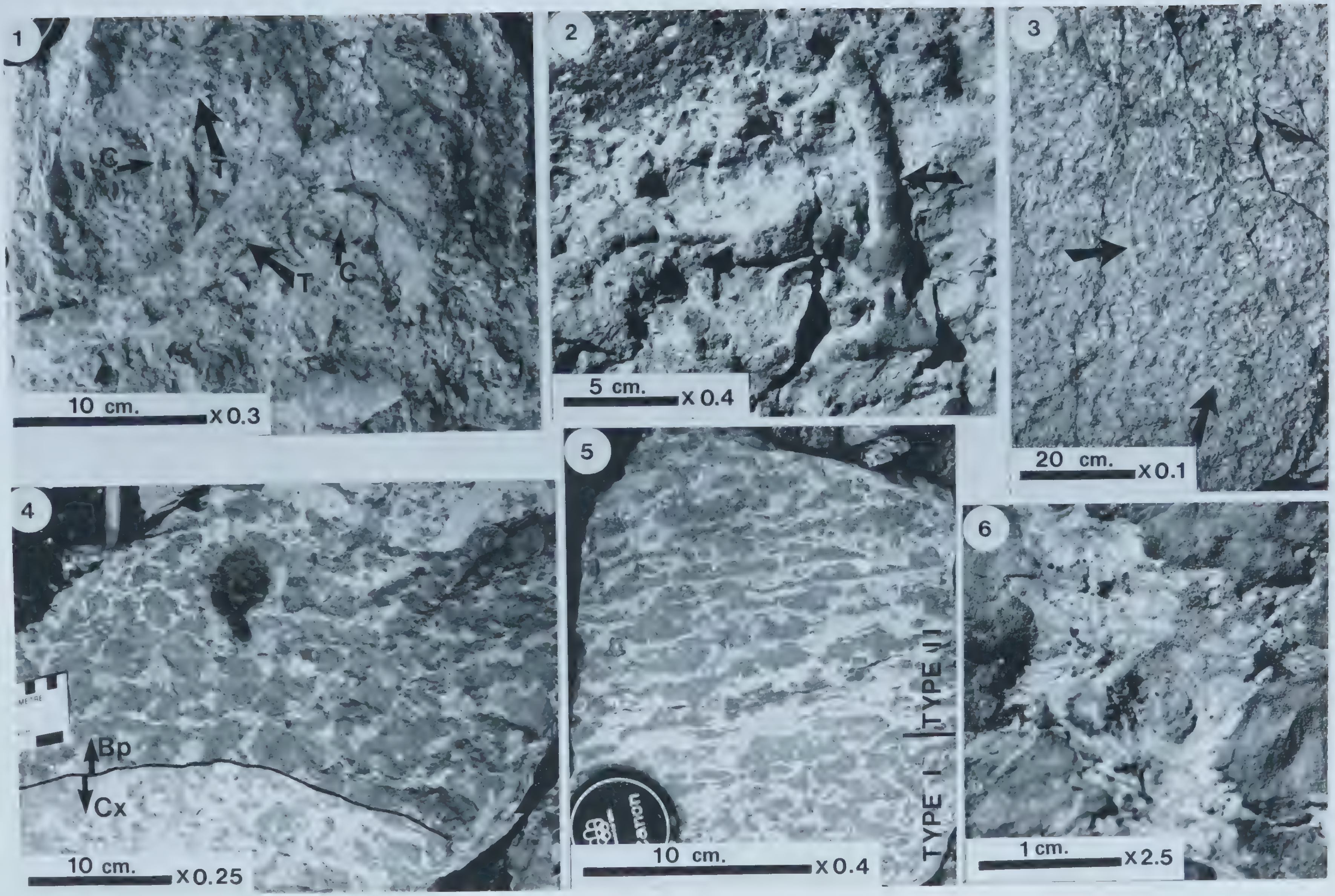
Figure 2. Field photograph of *Thalassinoides* burrows in a Type III mottled dolomitic limestone. The burrows show inflation at the points of branching and define a polygonal pattern on the bedding plane.
Unit K-69, Somerset Island Formation.

Figure 3. Field photograph of *Planolites* in a Type I mottled dolomitic limestone. Exposure is convex hyporelief.
Unit A-51, Douro Formation.

Figure 4. Field photograph of a *Thalassinoides* burrow system. The burrows (light grey) show a polygonal pattern on a bedding plane (Bp). The darker grey areas are the limestone lumps in a Type I mottled dolomitic limestone. Cx=cross-sectional view perpendicular to bedding.
Unit K-24, Douro Formation.

Figure 5. Field photograph of the vertical relationships of Types I and II mottled dolomitic limestone.
Unit HI-85, Douro Formation.

Figure 6. Field photograph of *Chondrites* burrows in the dolosiltite fill of a larger burrow in a Type I mottled dolomitic limestone.
Unit P-57, Douro Formation.


1
C
T
T
G
10 cm.
x 0.3
2
5 cm.
x 0.4
3
20 cm.
x 0.1
4
Bp
Cx
10 cm.
x 0.25
5
TYPE I
TYPE II
10 cm.
x 0.4
6
1 cm.
x 2.5

270

PLATE 20

Figure 1. Field photograph of the rubbly weathering of Types I and II mottled dolomitic limestone.
Units A-33 to A-38, Douro Formation.

Figure 2. Field photograph of the rubbly weathering of Types I and II mottled dolomitic limestone. The light grey areas are where the dolosiltite matrix has been replaced by chert.
Unit Q-24, Douro Formation.

Figure 3. Field photograph of the rubbly weathering of Types I and II mottled dolomitic limestone.
Unit A-38, Douro Formation.

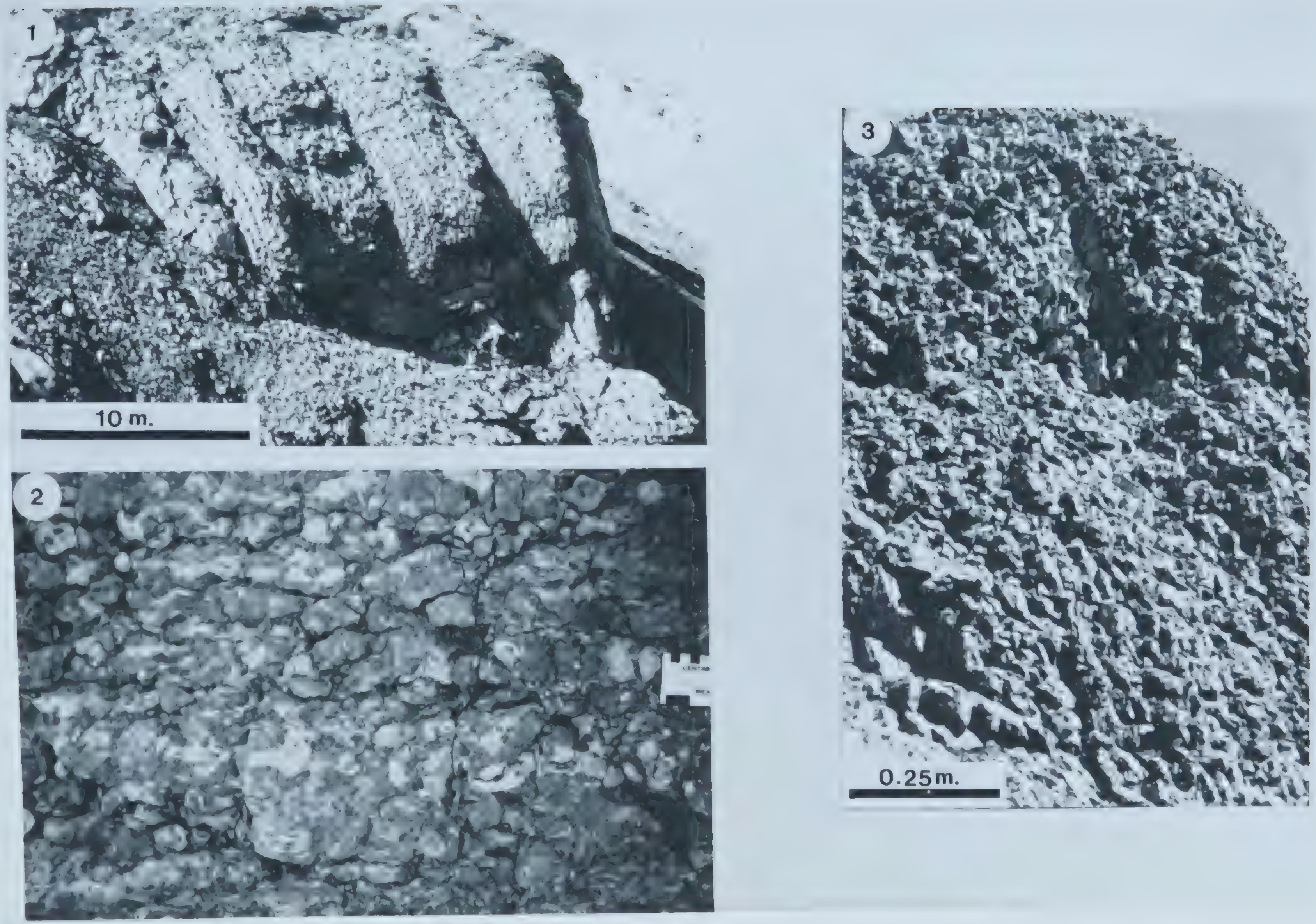

1
10 m.
2
3
0.25m.

PLATE 21

Figure 1. Photomicrograph of dolomitized mottled dolomitic limestone. The limestone has been replaced by xenotopic dolomite whereas the matrix retains its idiotopic texture. The xenotopic dolomite is a later diagenetic dolomite than the matrix.
Unit B-33, member C, Cape Storm Formation.

Figure 2. Photomicrograph of the limestone/dolosiltite interface.
Unit B-61, Douro Formation.

Figure 3. Photomicrograph of the idiotopic texture of the dolosiltite matrix. The dolomite rhombs are grain supported in a matrix of microcrystalline calcite. The limestone/dolosiltite interface is sharp.
Unit A-73, lower part of the Somerset Island Formation.

Figure 4. Photomicrograph of the limestone/dolosiltite interface.
Unit A-64, Douro Formation.

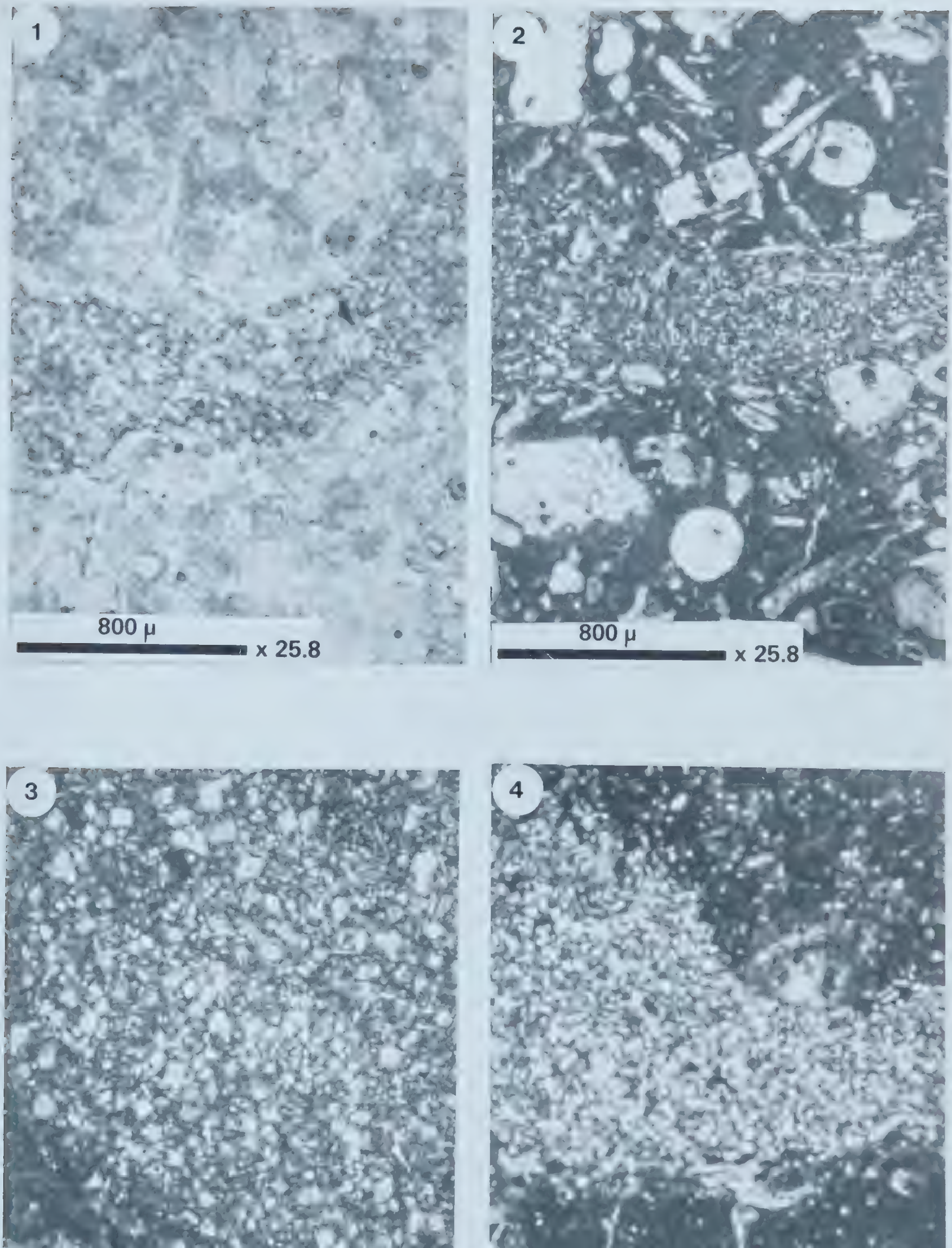

1
800 μ
x 25.8
2
800 μ
x 25.8
3
300 μ
x 65
4
800 μ
x 25.8

PLATE 22

Figure 1. Polished slab of a Type I mottled dolomitic limestone. Grey areas are the limestone lumps and the light areas are *Thalassinoides* burrows filled by the dolosiltite matrix. The color mottling in both lithologies are *Chondrites* burrows.
Unit G-65, Douro Formation.

Figure 2. Polished slab of a Type I mottled dolomitic limestone. The color mottling in the micrite is attributed to burrows of the preomission suite.
Unit B-70, Douro Formation.

Figure 3. Field photograph showing the irregular outline of the micrite lumps and layers in a Type II mottled dolomitic limestone. Photograph is perpendicular to bedding.
Unit K-32, lower part of the Somerset Island Formation.

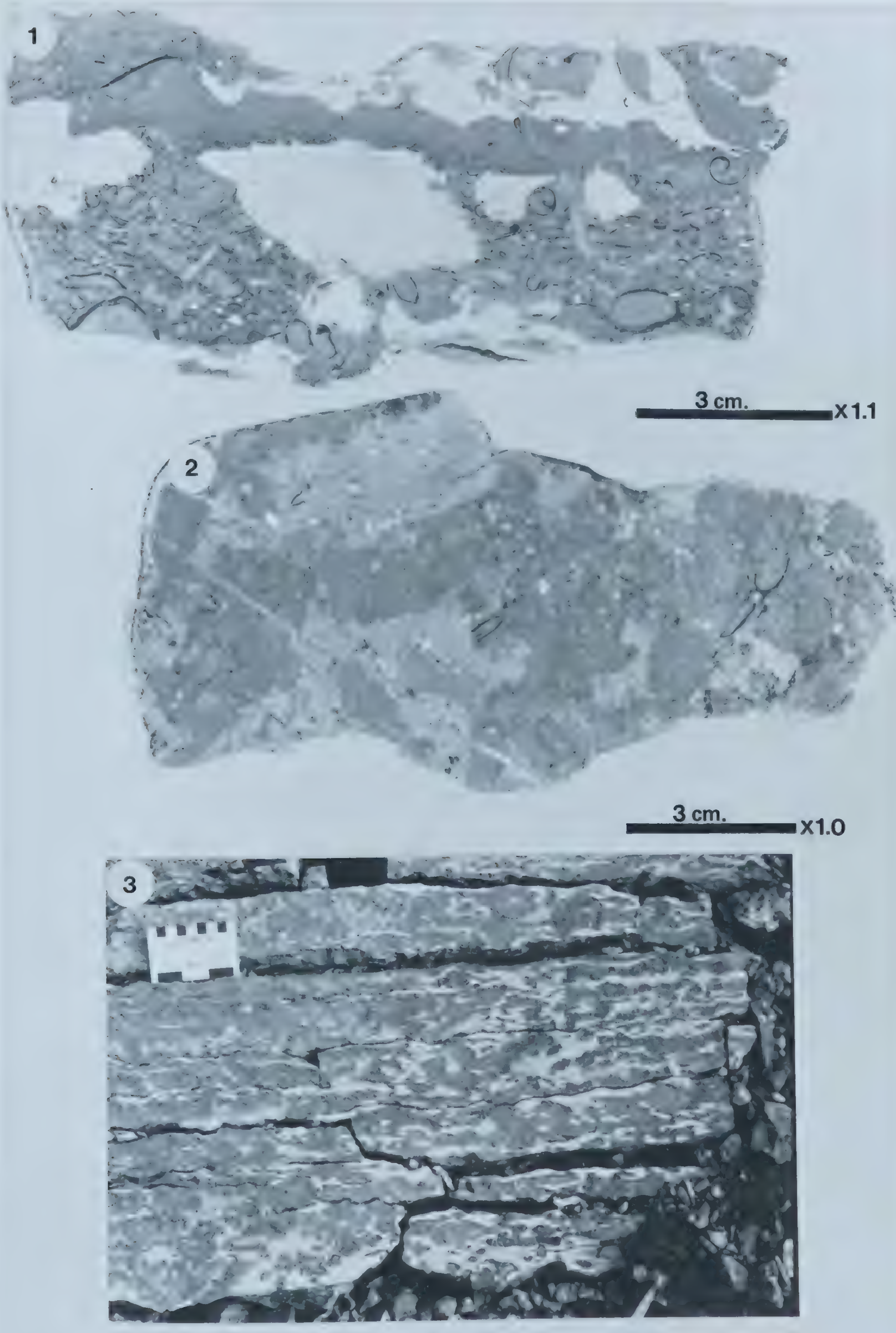

1
3 cm.
X1.1
2
3 cm.
X1.0
3

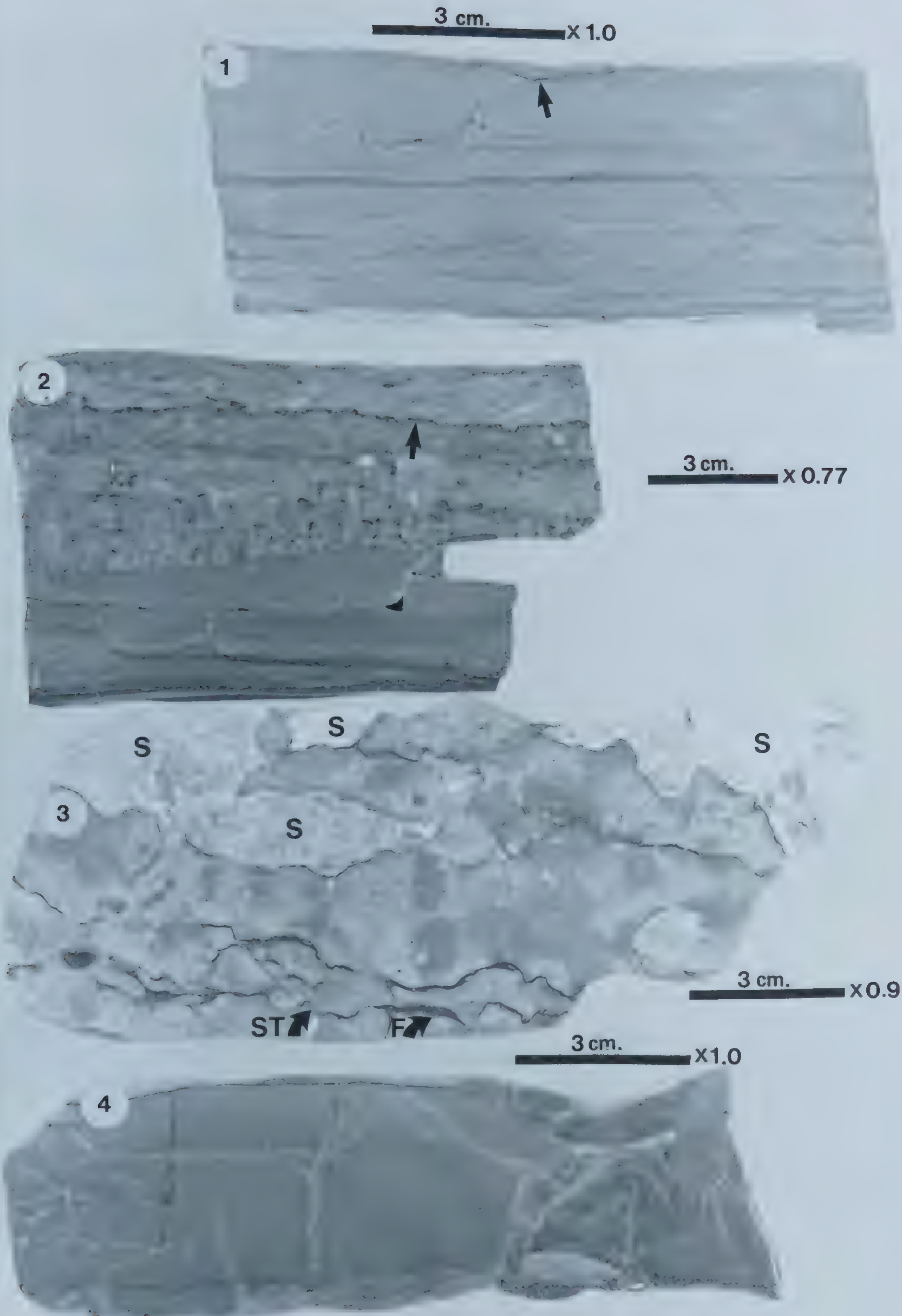
PLATE 23

Figure 1. Polished slab of a laminated dolostone. The arrow points to a mineralized hardground. Immediately below the arrow are small vertical burrows. The lower laminae in the slab show disruption by scour.
Unit HI-5, member A, Cape Storm Formation.

Figure 2. Polished slab of an interbedded limestone and dolostone. The arrow points to a mineralized hardground. A lesser developed hardground may occur at the base of the arrow. Bedding contacts in the lower portion of the slab show evidence of scour.
Unit L-40, member C, Cape Storm Formation.

Figure 3. Polished slab of mottled dolomitic limestone. A chronological succession of burrows is evident. Color mottling in the limestone lump is attributed to preomission bioturbation and the light grey areas are burrows of the omission suite. The burrows of the omission suite have been silicified (S). The lateral continuity of the burrows with stylolites (ST) and flaser structures (F) is apparent in this sample.
Unit L-67, Douro Formation.

Figure 4. Polished slab of a micrite lump. The top and bottom of the lump are marked by omission surfaces. The micrite shows evidence of early lithification, the fracture is infilled by the overlying sediment. The presence of intraclasts indicates some degree of lithification of the overlying sediment prior to fracturing. *Chondrites* burrows occur in the left side of the photograph.
Unit A-58, Douro Formation.



3 cm.
X 1.0
1
2
3 cm.
X 0.77
S
S
S
3
S
3 cm.
X0.9
ST
F
3 cm.
X1.0
4

PLATE 24

Figure 1. Photomicrograph of an erosion surface showing the truncation of a gastropod and an essentially planar surface with the overlying sediment.
Unit R-58, lower part of the Somerset Island Formation.

Figure 2. Photomicrograph of the erosion surface in Figure 1. Note the sharp and planar interface.
Unit R-58, lower part of the Somerset Island Formation.

Figure 3. Photomicrograph of an erosion surface truncating an oncolith. This is also shown on Plate 10, Figure 2.
Unit R-58, lower part of the Somerset Island Formation.

Figure 4. Photomicrograph of a hardground in a limestone lump in a Type II mottled dolomitic limestone. The hardground is relatively planar and marked by the presence of a mineralized surface and alignment of the clasts parallel to the hardground.
Unit M-62, lower part of the Somerset Island Formation.

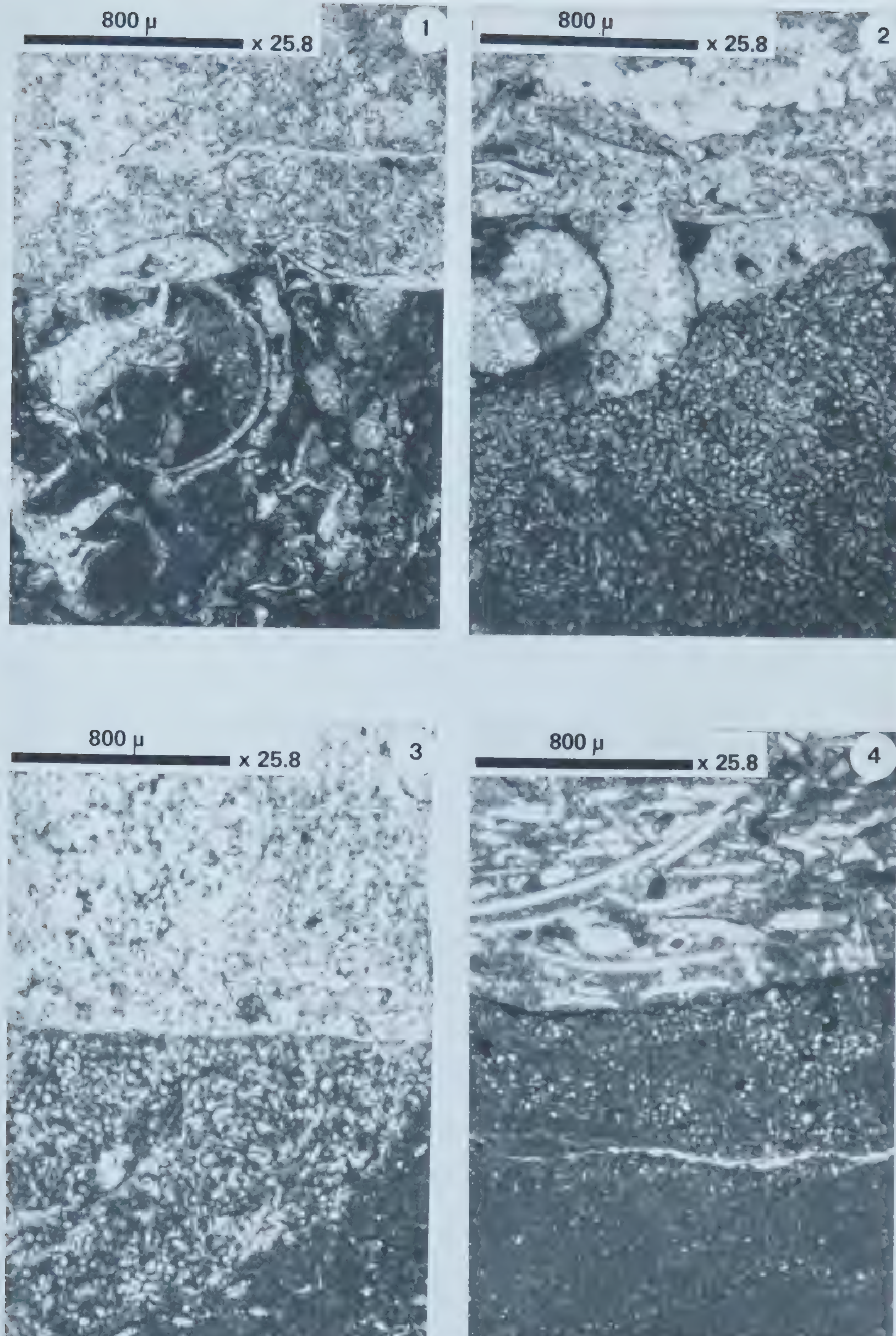

800 μ
x 25.8
1
800 μ
x 25.8
2
800 μ
x 25.8
3
800 μ
x 25.8
4

PLATE 25

Figure 1. Photomicrograph of *Chondrites* burrows of the preomission suite in a micrite lump.
Unit A-58, Douro Formation.

Figure 2. Field photograph of *Palaeophycus* burrows in a Type I mottled dolomitic limestone. Exposure is convex hyporelief.
Unit N-116, Somerset Island Formation.

Figure 3. Photomicrograph of a *Chondrites* burrow of the preomission suite of a micrite lump. The burrow-fill has been partially replaced by neomorphic sparite.
Unit G-72, Douro Formation.

Figure 4. Field photograph of *Thalassinoides* burrows in a Type I or II mottled dolomitic limestone. Exposure is convex hyporelief.
Unit L-59, Douro Formation.

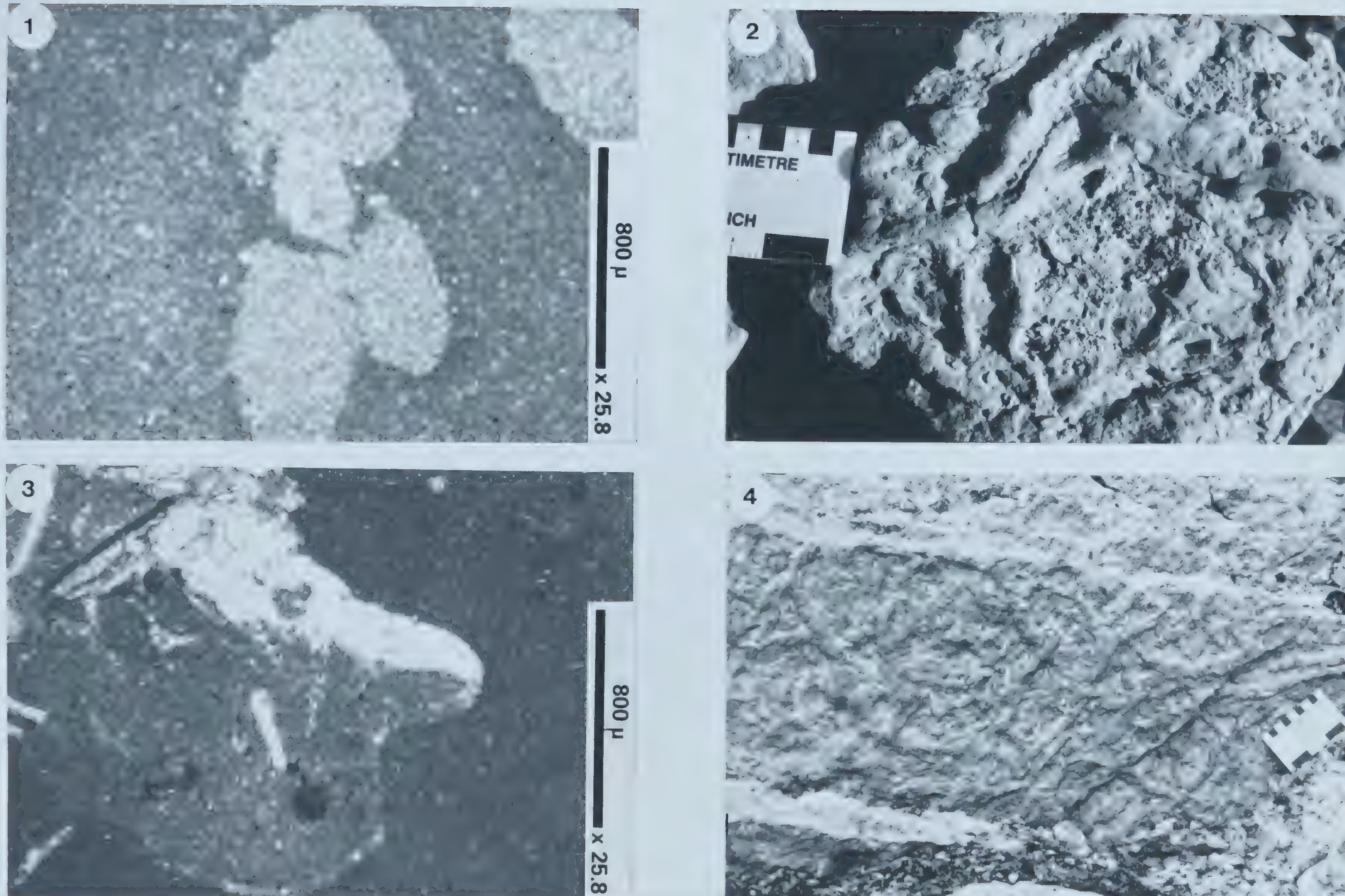

1
800 µ
x 25.8
2
TIMETRE
ICH
3
800 µ
x 25.8
4

PLATE 26

Figure 1. Field photograph of *Phycodes* burrows, exposure is convex hyporelief. Burrow-fill is a dolosiltite. Unit HI-118, Somerset Island Formation.

Figure 2. Field photograph of *Chondrites* in a dolosiltite matrix of a Type III mottled dolomitic limestone. Burrows are of the postomission suite. Unit Q-57, Somerset Island Formation.

Figure 3. Field photograph showing the radial tendency of *Chondrites*. Burrows occur in the postomission suite in the dolosiltite matrix of a Type III mottled dolomitic limestone. Unit N-90, Somerset Island Formation.

Figure 4. Field photograph of *Chondrites* in a Type I or II mottled dolomitic limestone. Unit M-36, Douro Formation.

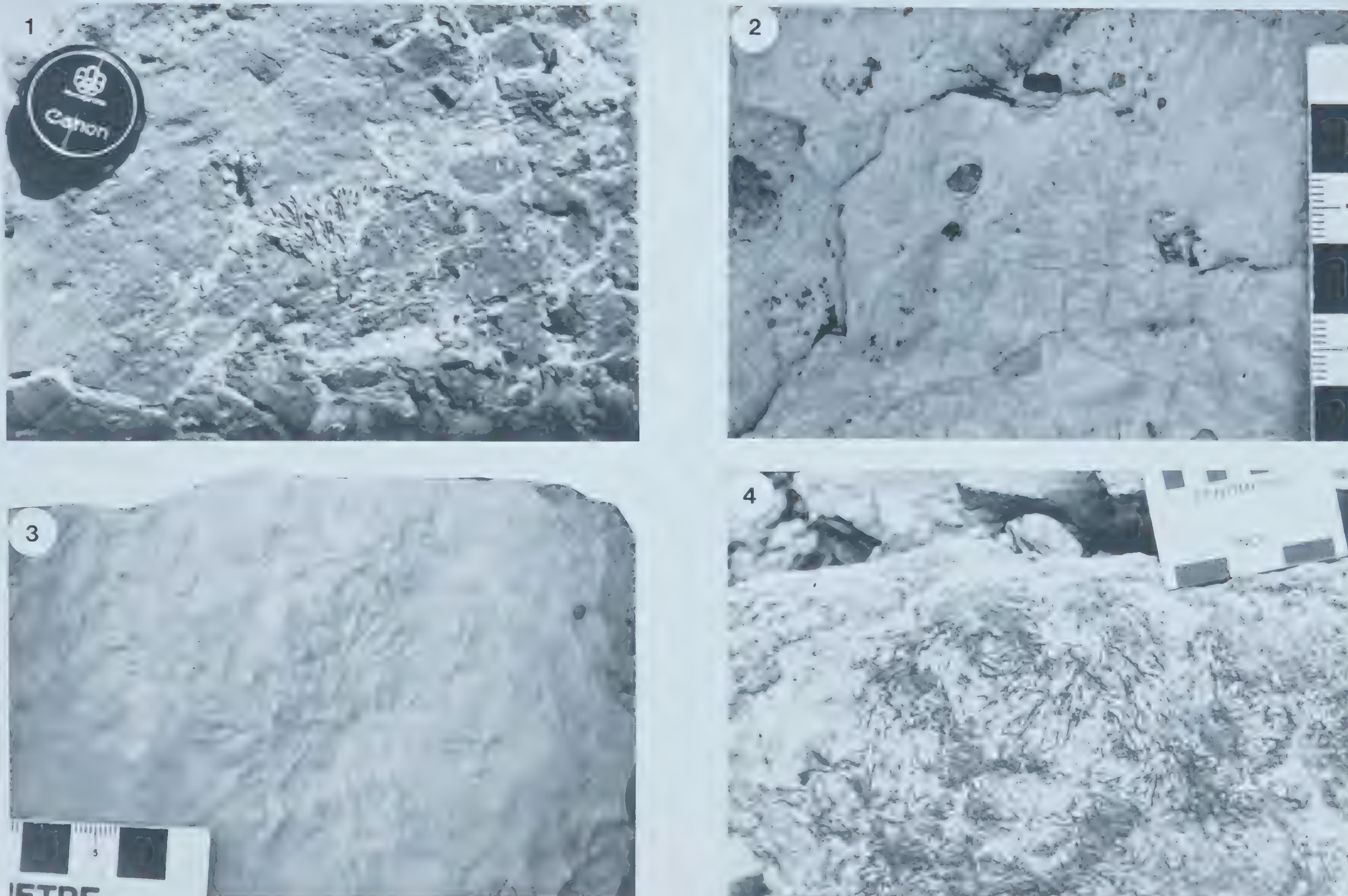

1
2
3
4
Canon

PLATE 27

Figure 1. Field photograph of *Chondrites* burrows in the dolosiltite matrix of a Type III mottled dolomitic limestone. Burrows are of the postomission suite. Unit N-4, member C, Cape Storm Formation.

Figure 2. Field photograph of *Gordia* burrows in a massive dolosiltite. Exposure is convex hyporelief. Unit L-17, member C, Cape Storm Formation.

Figure 3. Photomicrograph of a vertical burrow in a laminated dolosiltite. Unit HI-1, member A, Cape Storm Formation.

Figure 4. Field photograph of *Chondrites* burrows of the omission suite in a Type III mottled dolomitic limestone. Unit N-14, member C, Cape Storm Formation.

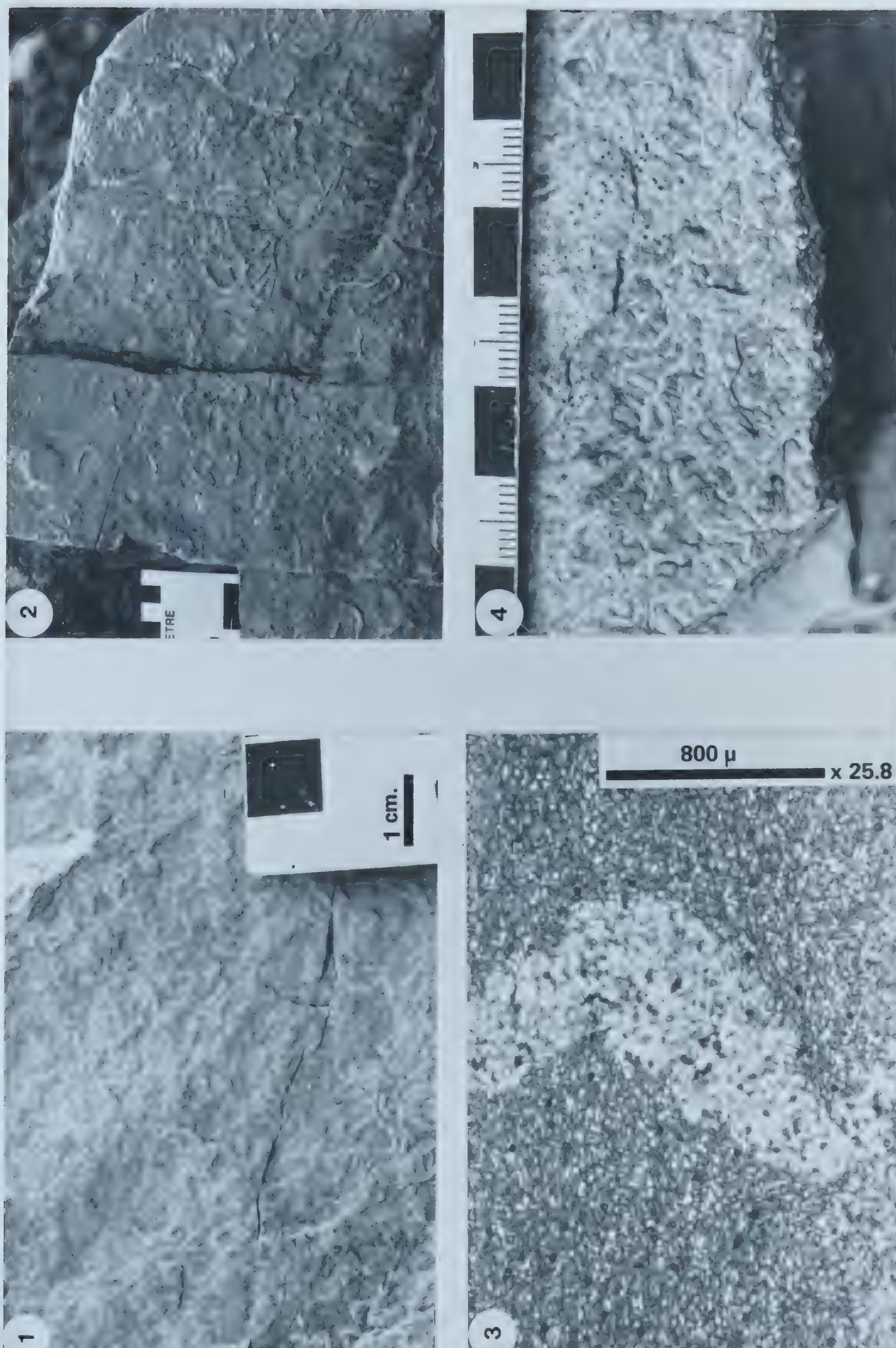

1 cm.
800 µ
x 25.8
1
2
3
4

PLATE 28

Figure 1. Photomicrograph of a crinoid ghost fabric in a dolomitized mottled dolomitic limestone. The ossicle retains its unicrystal structure while in a xenotopic matrix. Plane-polarized light.
Unit G-61, Douro Formation.

Figure 2. As in Figure 1., using cross-polarized light.
Unit G-61, Douro Formation.

Figure 3. Photomicrograph of a ghost fabric in a partially dolomitized limestone.
Unit G-95, Somerset Island Formation.

Figure 4. Photomicrograph of ghost fabrics in a dolostone. Original fossil was possibly an ostracod.
Unit D-77, Somerset Island Formation.

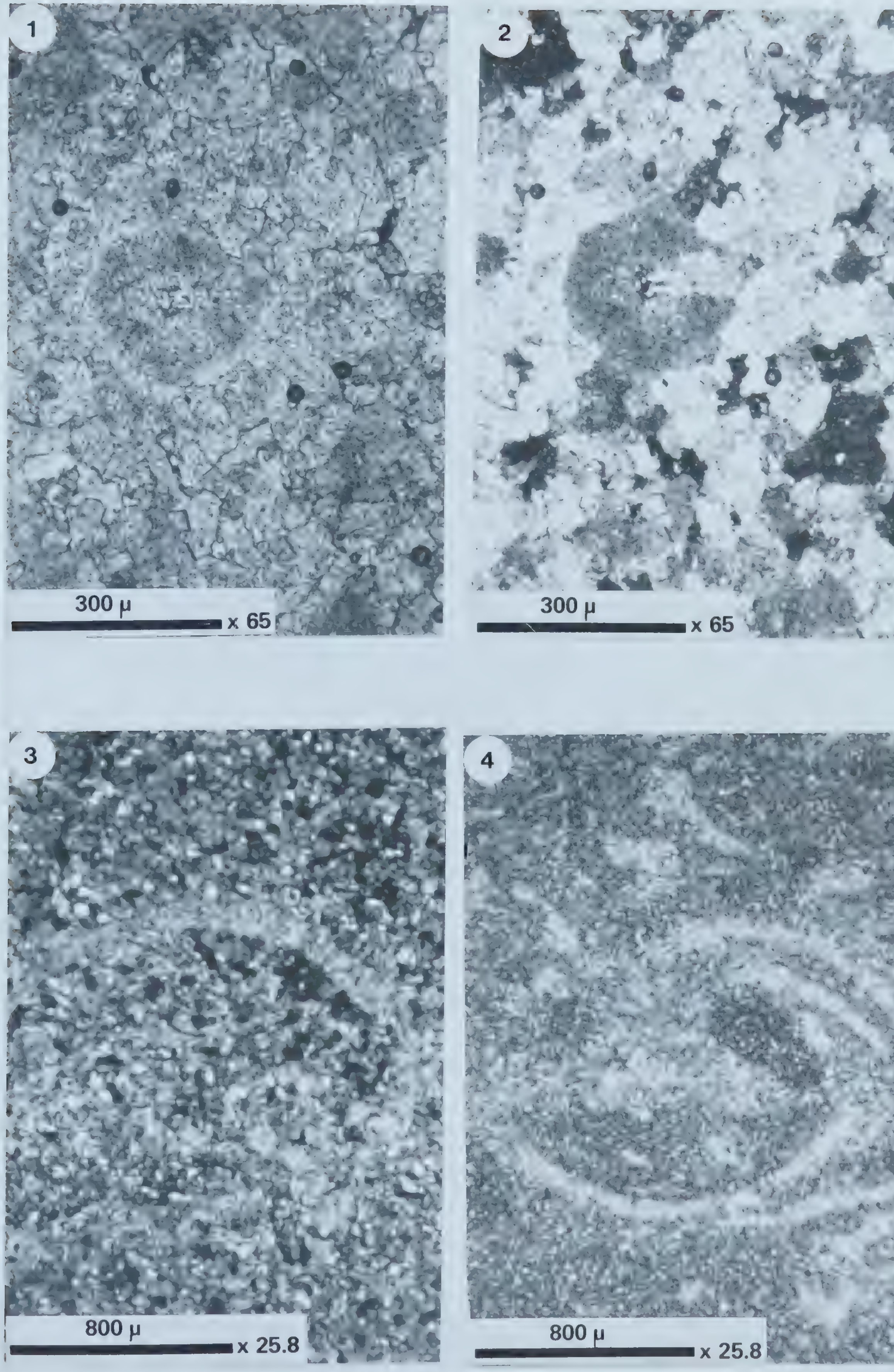

1
300 µ
x 65
2
300 µ
x 65
3
800 µ
x 25.8
4
800 µ
x 25.8

PLATE 29

Figure 1. Photomicrograph of a polycrystalline calcite rhomb (dedolomitization). Recognition of the dedolomite is by the relict outline of the rhomb. Plane-polarized light. Unit A-72, basal Somerset Island Formation.

Figure 2. As in Figure 1., using cross-polarized light. The rhomb is more apparent under cross-polarized light. Crystal faces of the rhomb are corroded and the calcite mimics the limestone texture.
Unit A-72, basal Somerset Island Formation.

Figure 3. Photomicrograph of a zoned dolomite.
Plane-polarized light.
Unit A-69, upper Douro Formation.

Figure 4. As in Figure 3., using cross-polarized light.
Outer rim of calcite is in optical continuity with the calcite within the rhomb.
Unit A-69, upper Douro Formation.

Figure 5. Photomicrograph of a dedolomite. The dedolomite mimics the texture of the limestone.
Unit A-71, basal Somerset Island Formation.

Figure 6. As in Figure 5., using cross-polarized light. The rhomb is accentuated under cross-polarized light.
Centripetal replacement of dolomite by calcite.
Unit A-71, basal Somerset Island Formation.

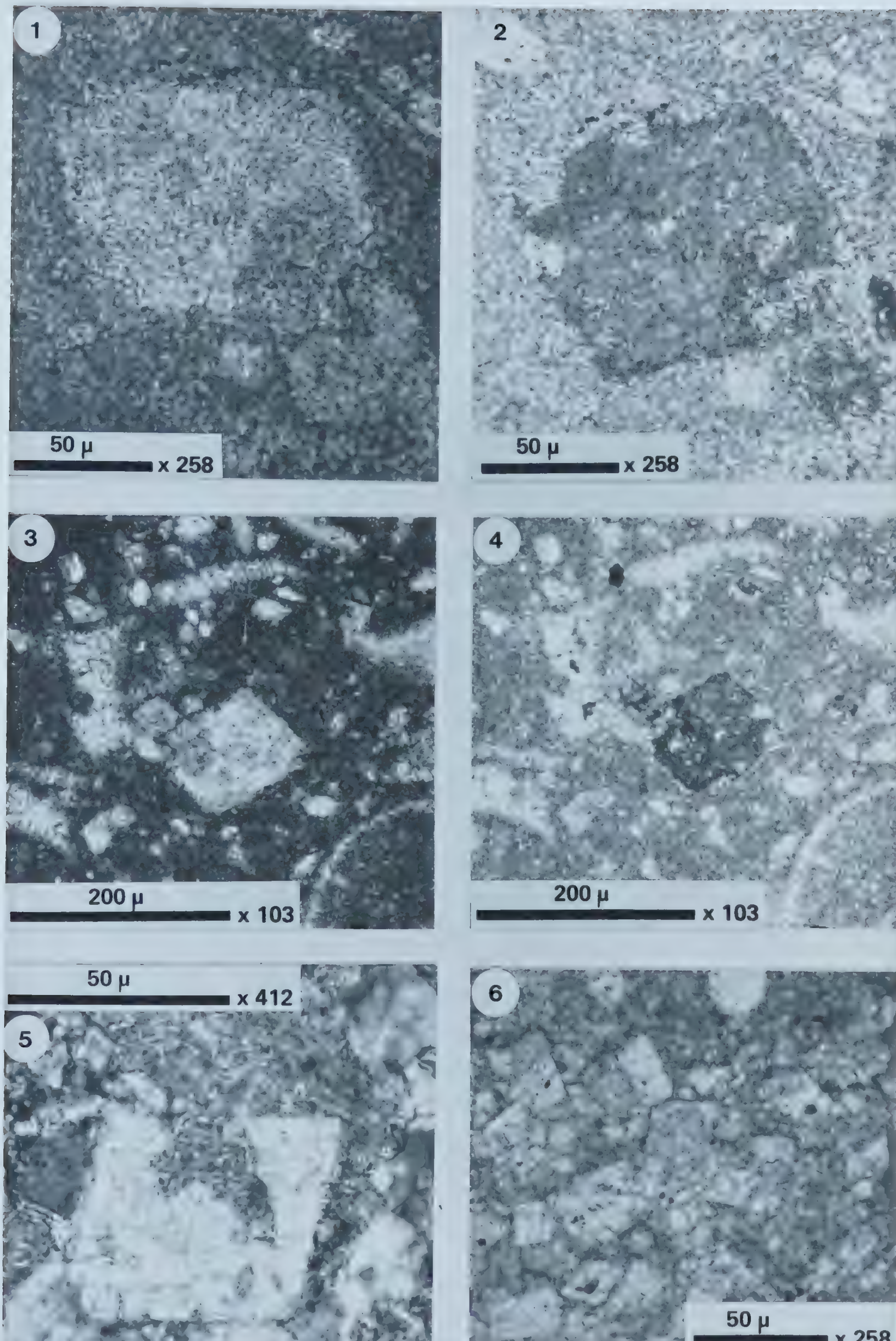

1
50 μ
x 258
2
50 μ
x 258
3
200 μ
x 103
4
200 μ
x 103
50 μ
x 412
5
6
50 μ
x 258

PLATE 30

Figure 1. Photomicrograph of a dolomite rhomb totally replaced by calcite. In the lower right corner of the photograph, the rhomb shows further corrosion. Plane-ploarized light.
Unit A-71, basal Somerset Island Formation.

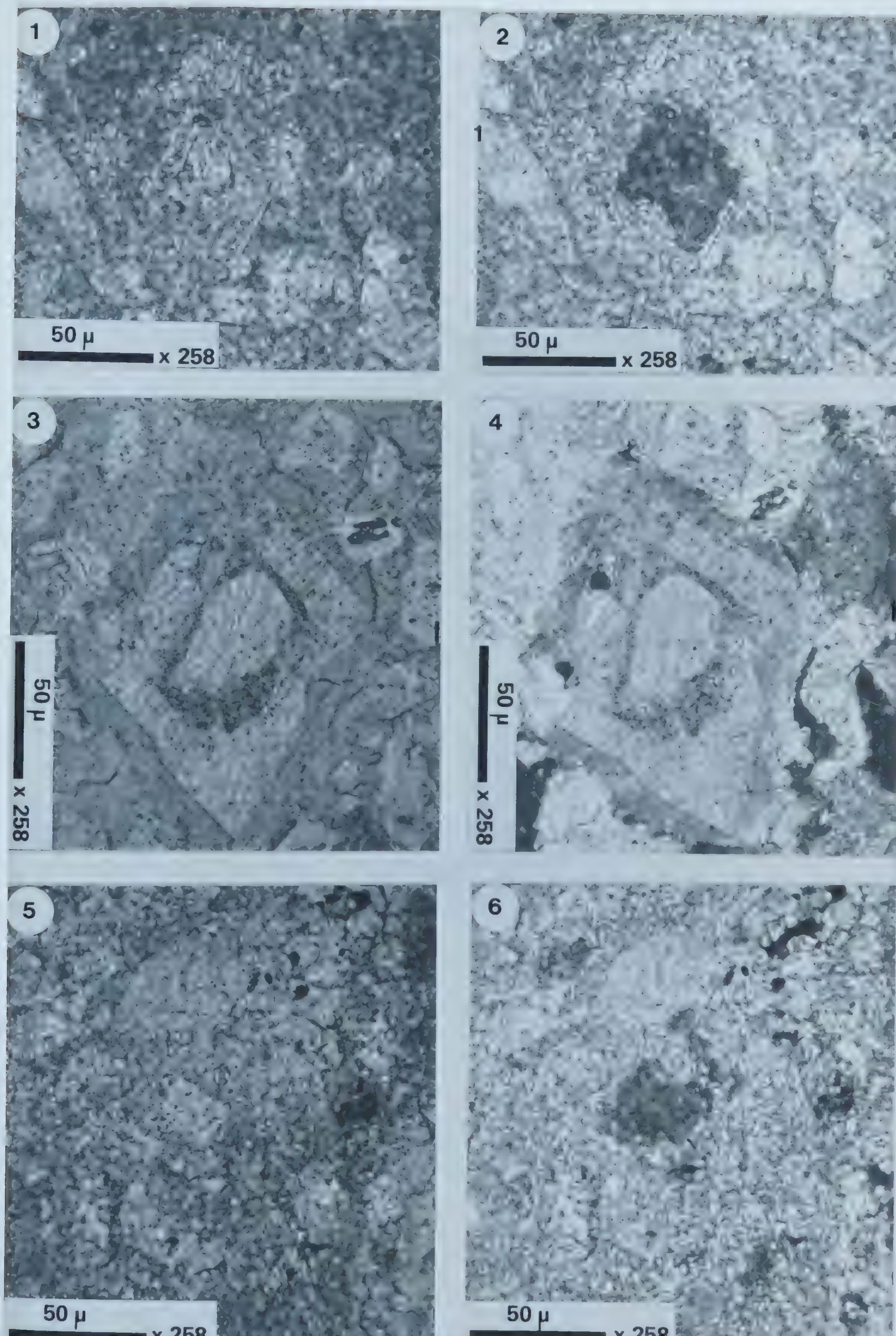
Figure 2. As in Figure 1., using cross-polarized light. The corrosion in the lower right corner of the photograph shows the calcite mimicing the limestone texture.
Unit A-71, basal Somerset Island Formation.

Figure 3. Photomicrograph of dedolomitization in a micrite lump. Several smaller rhombs are evident in the upper right corner of the photograph. Centrifugal replacement of dolomite by calcite. Plane-polarized light.
Unit A-71, basal Somerset Island Formation.

Figure 4. As in Figure 3., using cross-polarized light. The smaller rhombs show the calcite mimicing the limestone texture.
Unit A-71, basal Somerset Island Formation.

Figure 5. Photomicrograph of a dolomite rhomb showing a calcite embayment. The calcite mimics the texture of the limestone. Plane-polarized light.
Unit N-56, basal Somerset Island Formation.

Figure 6. Photomicrograph of dedolomites. Most of the rhombs show partial dedolomitization. Some of the rhombs in the lower left corner of the photograph may show dissolution to form rhombohedral pores.
Unit N-56, basal Somerset Island Formation.



1
50 μ
x 258
2
1
50 μ
x 258
3
50 μ
x 258
4
50 μ
x 258
5
50 μ
x 258
6
50 μ
x 258

PLATE 31

Figure 1. Negative print of a cross-laminated dolosiltite. Unit B-19, member A, Cape Storm Formation.

Figure 2. Negative print of a quartzose dolosiltite. Note the laminated intraclast and the vertical burrow. Unit A-2, member A, Cape Storm Formation.

Figure 3. Photomicrograph of a graded dolosiltite with a microfault displacing the laminae. Unit L-15, member A, Cape Storm Formation.

Figure 4. Negative print of graded bedding in a dolosiltite. Note the vertical burrow. Unit HI-1, member A, Cape Storm Formation.

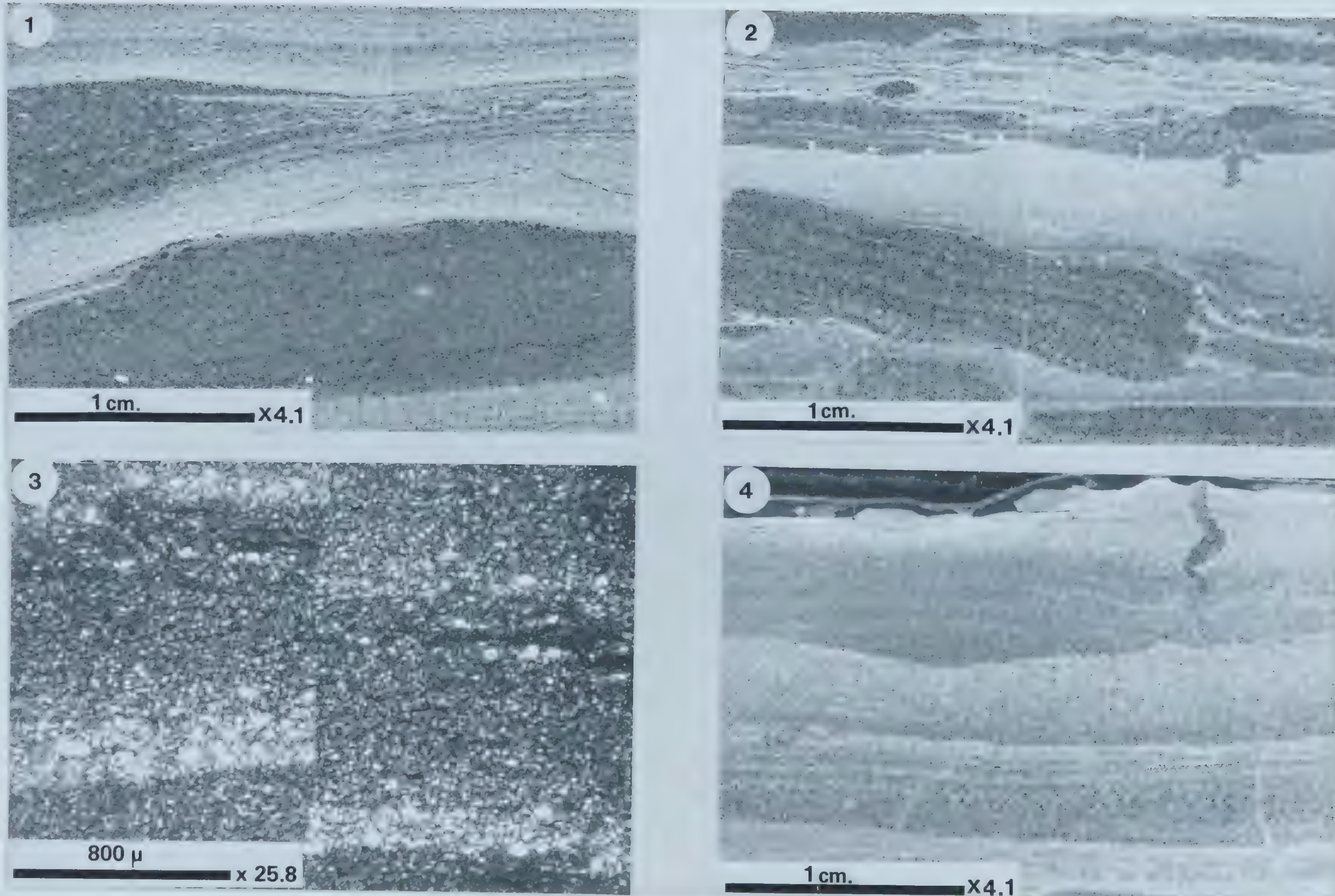

1
1 cm.
X4.1
2
1 cm.
X4.1
3
800 μ
x 25.8
4
1 cm.
X4.1

PLATE 32

Figure 1. Field photograph of ripple marks in a dolomitic quartz siltstone.
Unit Q-109, upper Somerset Island Formation.

Figure 2. Field photograph of ripple marks in a dolomitic quartz arenite.
Unit Q-110, upper Somerset Island Formation.

Figure 3. Field photograph of ripple marks in a quartzose dolosiltite. Note the burrow on the ripple crest.
Unit P-19, member A, Cape Storm Formation.

Figure 4. Modern day analogue of ripple marks and burrows during low tide on a clastic tidal flat, Cresent Beach, British Columbia.

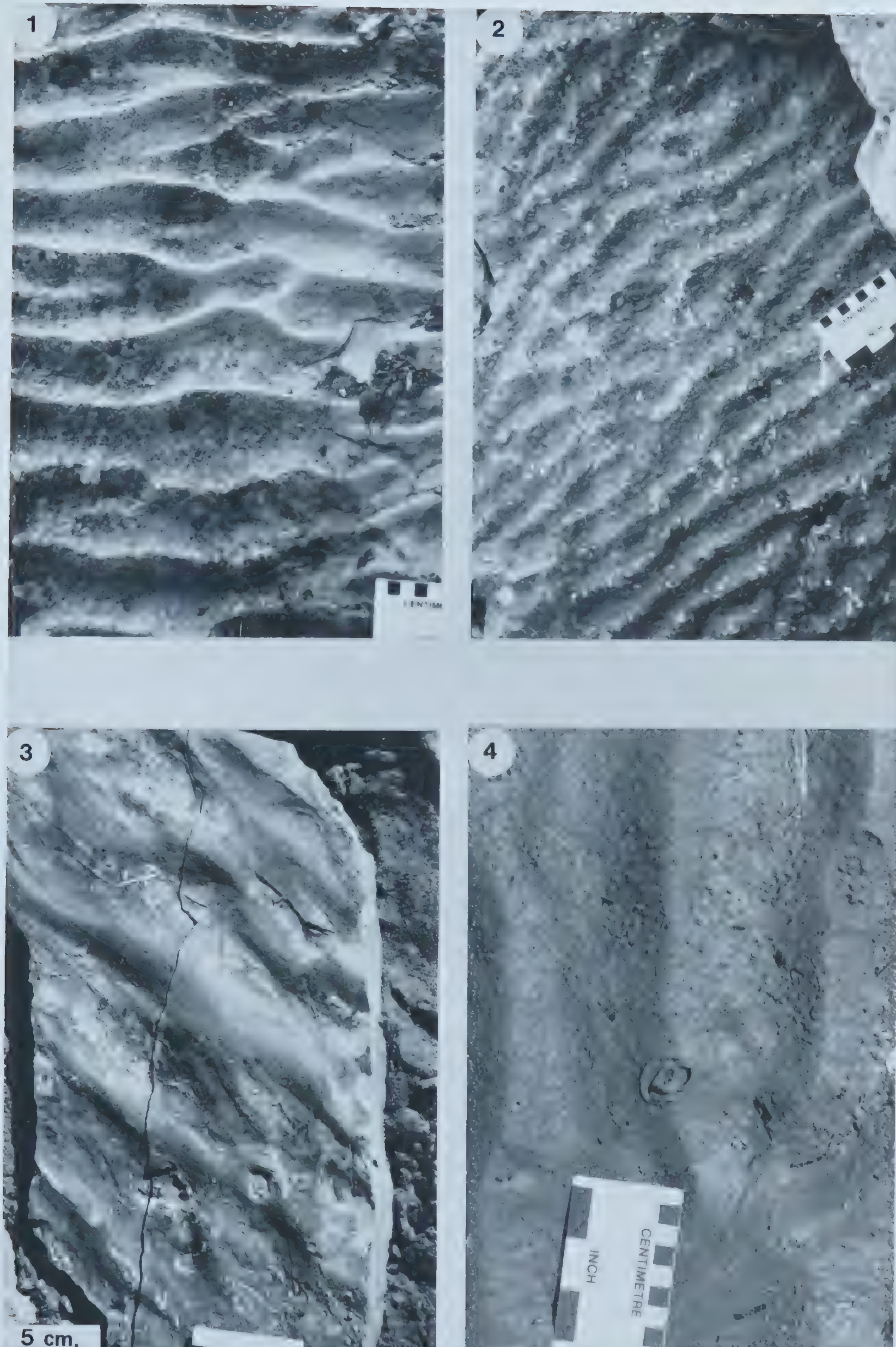

1
2
3
5 cm.
4
CENTIMETRE
INCH

PLATE 33

Figure 1. Photomicrograph of sand-sized detrital dolomite grains. The original grain shows impurities and a rimming of opaques along the grain margin. Plane-polarized light.
Unit B-6, member A, Cape Storm Formation.

Figure 2. As in Figure 1, using cross-polarized light.
UNit B-6, member A, Cape Storm Formation.

Figure 3. Photomicrograph of sand-sized detrital dolomite grains. Sample is a dolosiltite to dolarenite from the interarea between hemispherical stromatolites.
Plane-polarized light.
Unit B-19, member B, Cape Storm Formation.

Figure 4. As in Figure 3, using cross-polarized light.
Unit B-19, member B, Cape Storm Formation.

Figure 5. Photomicrograph of a sand-sized detrital dolomite grain. Note the rhombohedral form of the overgrowth. Original grains is apparent by a thin line of opaques along the grain margin. Plane-polarized light.
UNit B-96, lower part of the Somerset Island Formation.

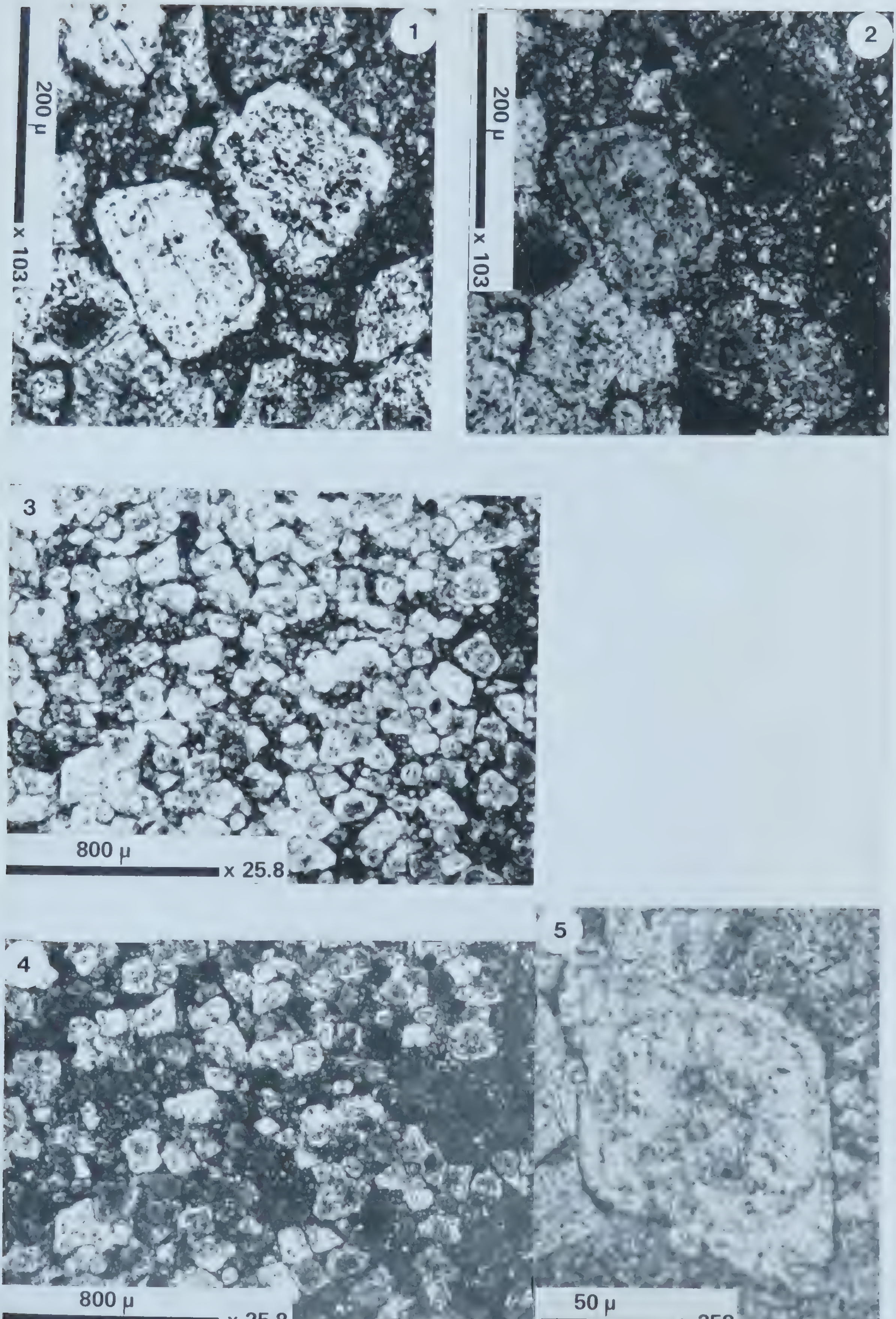

1
200 μ
x 103
2
200 μ
x 103
3
800 μ
x 25.8
4
800 μ
x 25.8
5
50 μ
x 258

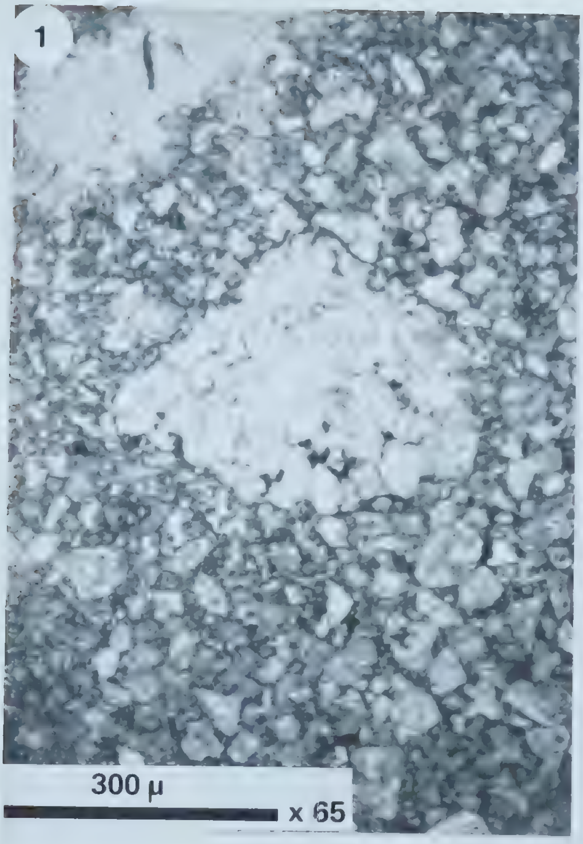
PLATE 34

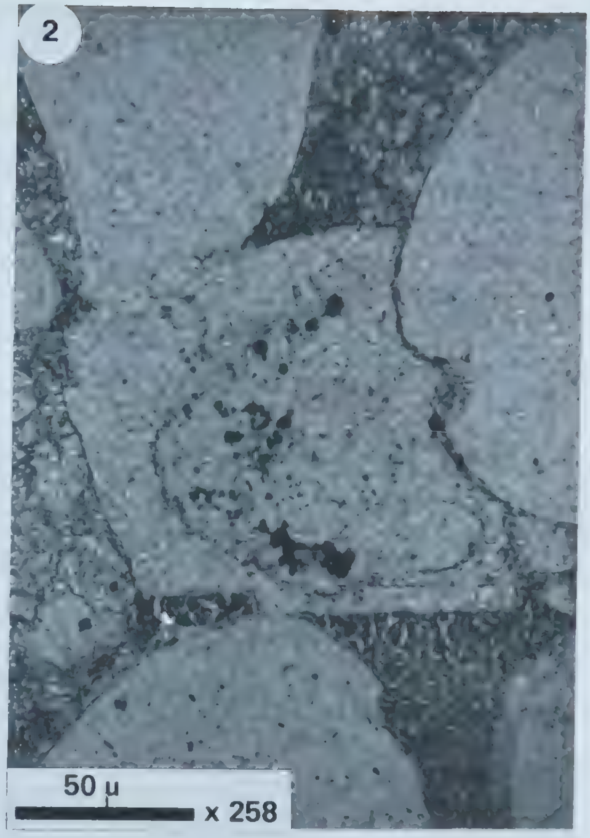
Figure 1. Photomicrograph of a polycrystalline detrital dolomite grain. Plane-polarized light.
Unit N-157, upper part of the Somerset Island Formation.

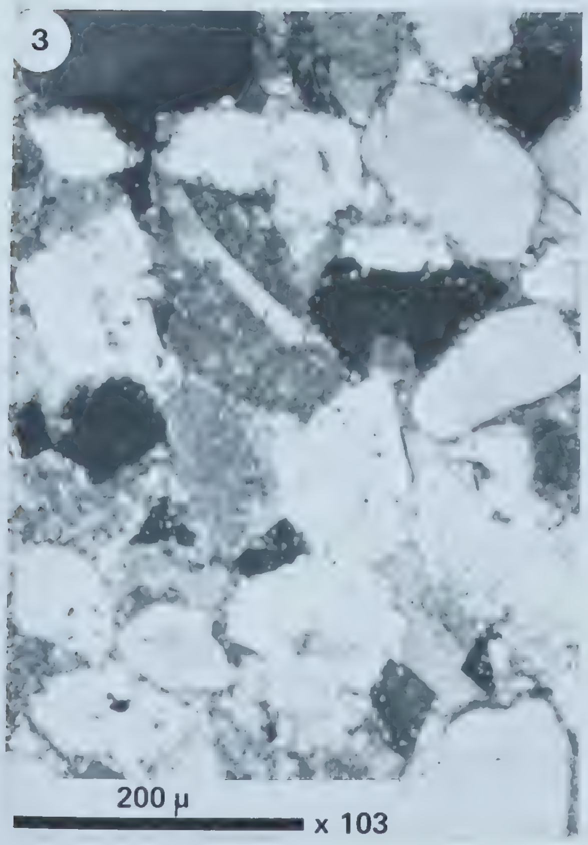
Figure 2. Photomicrograph of a detrital dolomite grain surrounded by quartz grains. The grain and its overgrowth show a pressure contact with a quartz grain. Original grain margin is marked by a thin line of opaques. Cement is calcite. Plane-polarized light.
Unit K-76, Somerset Island Formation.

Figure 3. Photomicrograph of detrital dolomite, micrite and quartz grains. The dolomite grains show well developed syntaxial overgrowths whereas the micrite and quartz grains do not. Cross-polarized light.
Unit K-71, Somerset Island Formation.

Figure 4. As in Figure 3, using plane-polarized light.
Unit K-71, Somerset Island Formation.


1
300 μ
x 65


2
50 μ
x 258


3
200 μ
x 103


4
200 μ
x 103

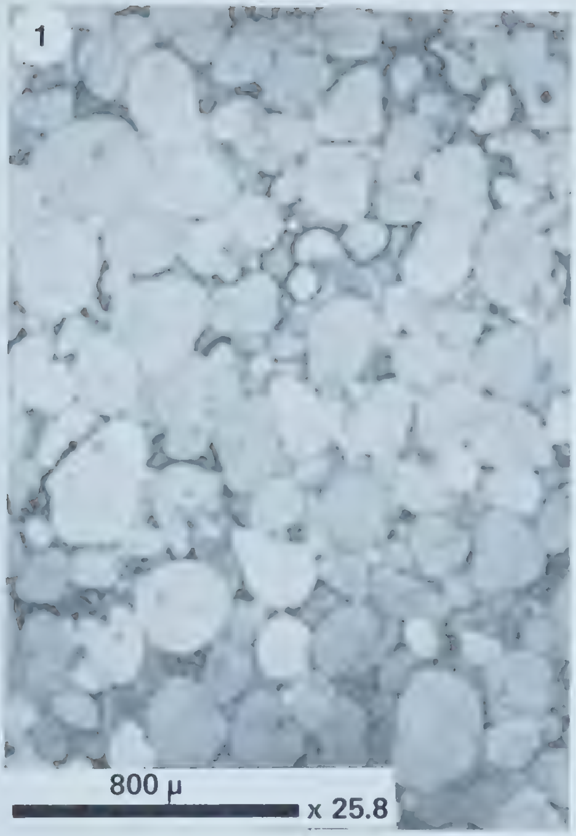
PLATE 35

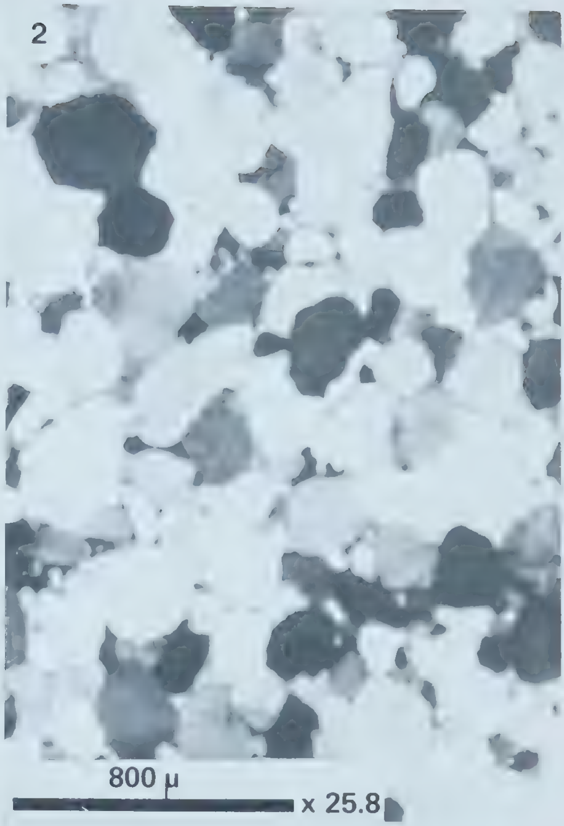
Figure 1. Photomicrograph of a calcareous cemented quartz arenite at the Douro Formation-Somerset Island Formation contact. Note the roundness and degree of sorting of the quartz grains. Plane-polarized light.
Unit K-27, basal Somerset Island Formation.

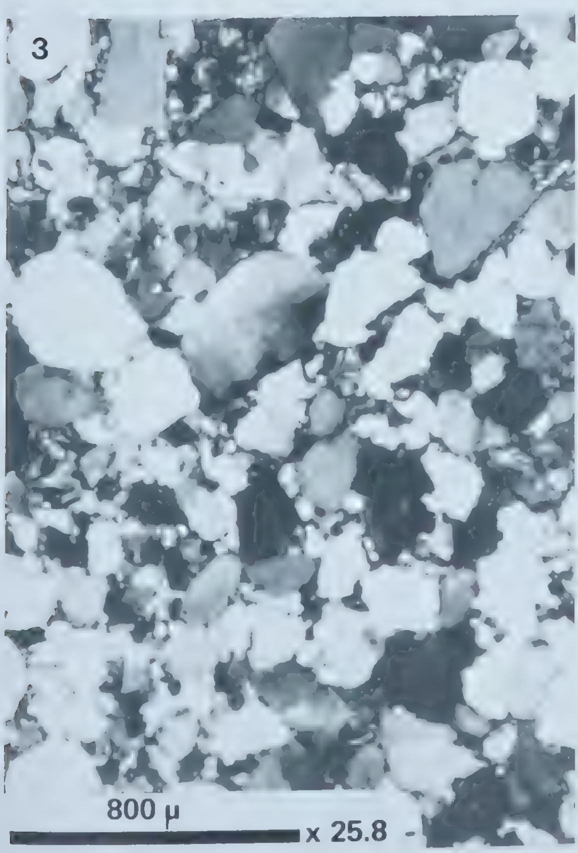
Figure 2. As in Figure 1, using cross-polarized light.
Unit K-27, basal Somerset Island Formation.

Figure 3. Photomicrograph of a sublitharenite in the upper part of the Somerset Island Formation. Note the presence of stressed and polycrystalline grains and the lesser degree of sorting than in Figures 1 and 2.
Cross-polarized light.
Unit K-77, upper part of the Somerset Island Formation.

Figure 4. Field photograph of the Douro Formation-Somerset Island Formation contact. The Douro Formation is prominent whereas the Somerset Island Formation weathers more recessive. Boundary is between units Q-32 and Q-33.
UNits Q-30 to Q-42, Douro and Somerset Island formations.


1
800 μ
x 25.8


2
800 μ
x 25.8


3
800 μ
x 25.8

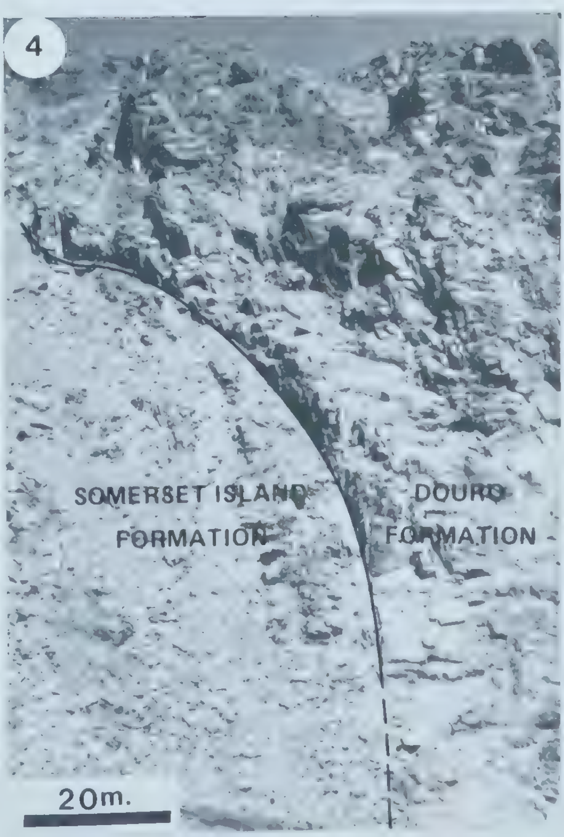
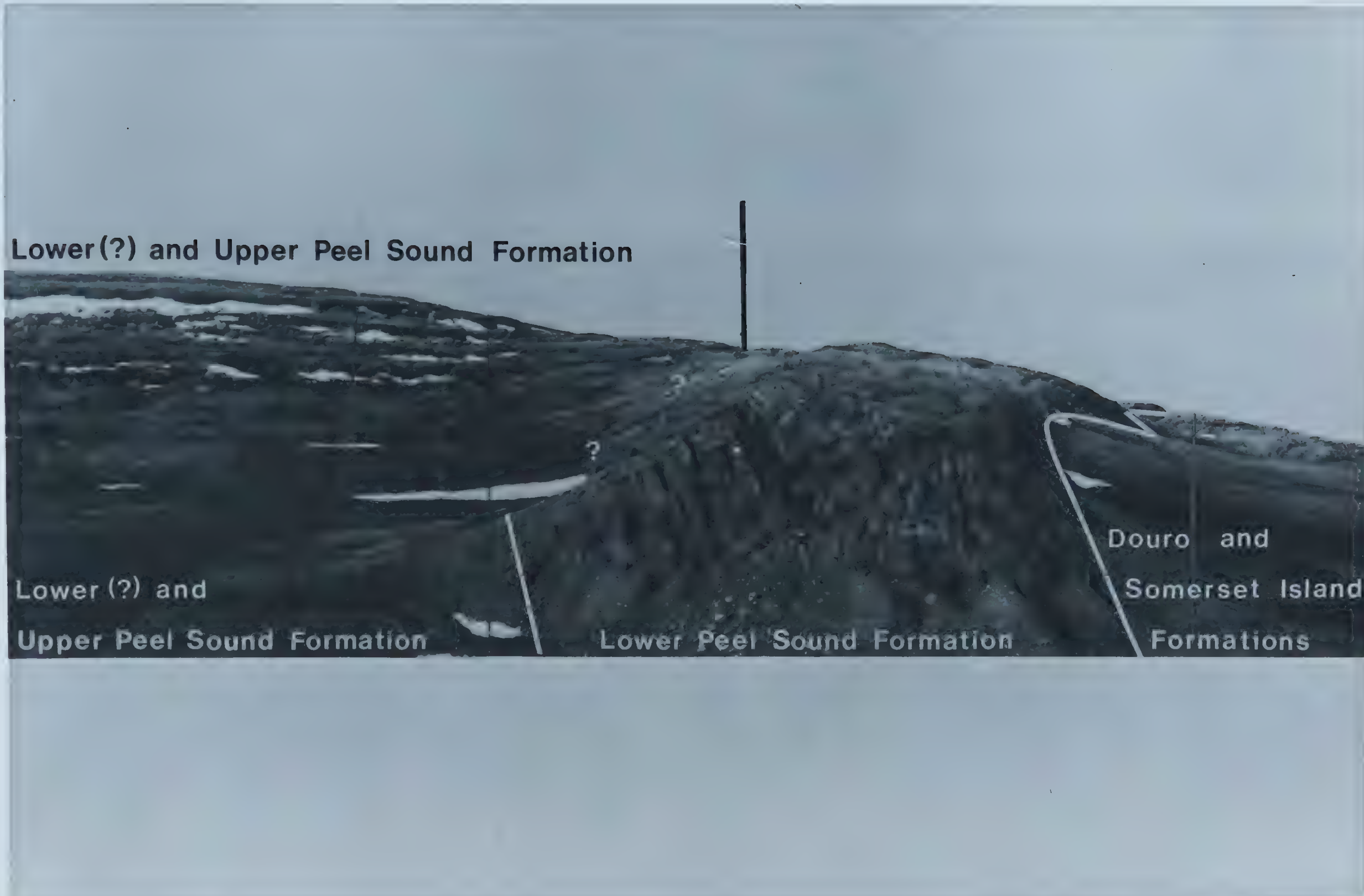

4
SOMERSET ISLAND
FORMATION
DOURO
FORMATION
20m.

PLATE 36

Figure 1. Structural style at Le Feurve Inlet. The overturned strata mark the western limit of the Cornwallis Fold Belt and the horizontal strata mark the essentially undeformed strata of the Arctic Lowlands. Coastal section in the photograph is approximately 2 km south of Section G.



Lower(?) and Upper Peel Sound Formation
Lower (?) and
Upper Peel Sound Formation
Lower Peel Sound Formation
Douro and
Somerset Island
Formations

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XI. APPENDIX I, STRATIGRAPHIC SECTIONS

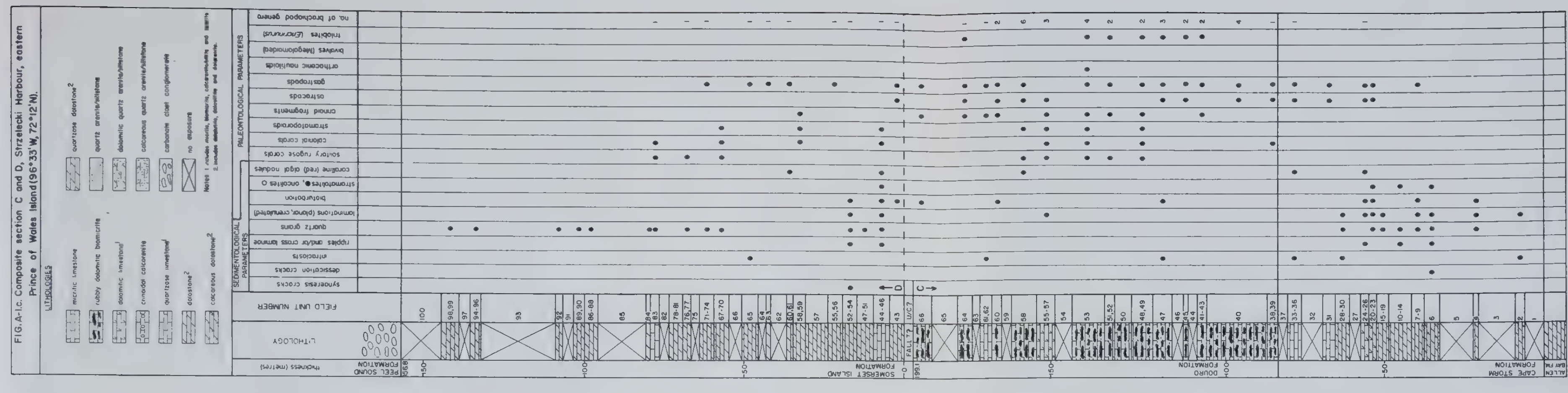


FIG. A-1c. Composite section C and D, Strzelecki Harbour, eastern Prince of Wales Island (96°33'W, 72°12'N).

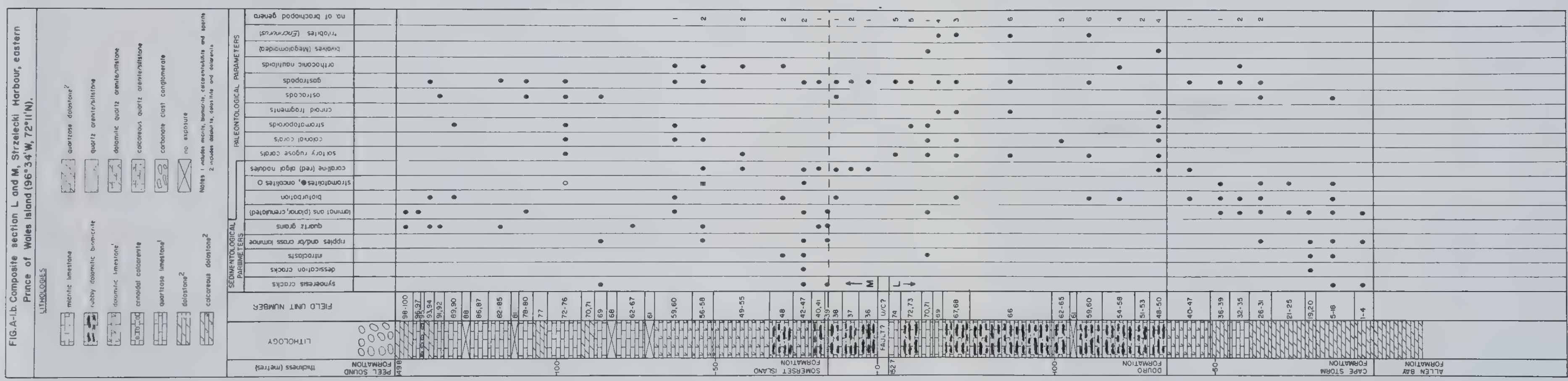


FIG. A-1b. Composite section L and M, Strzelecki Harbour, eastern Prince of Wales Island (96°34'W, 72°11'N).

FIG. A-1a. Section N, "Wilson's Gorge", south Strzelecki Harbour, eastern Prince of Wales Island (96°30'W, 72°10'N).

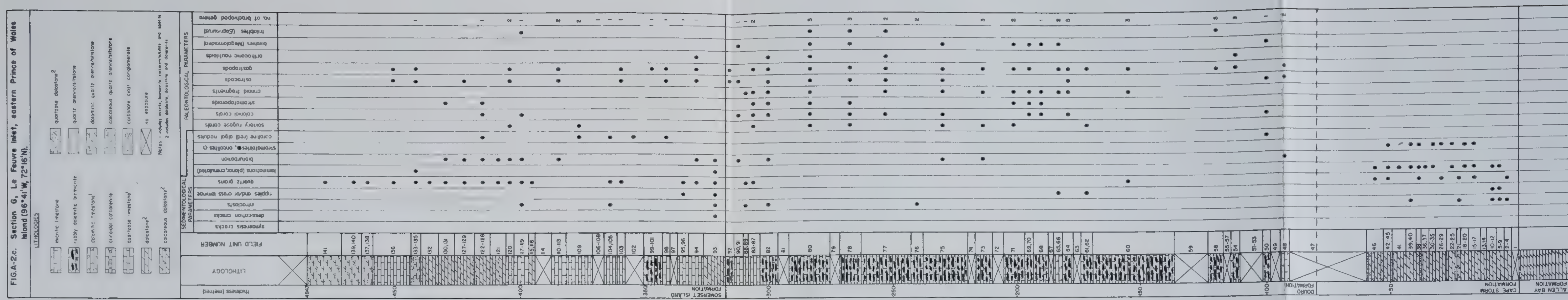


FIG. A-2c. Section G, Le Feuvre Inlet, eastern Prince of Wales Island (96°41'W, 72°16'N).

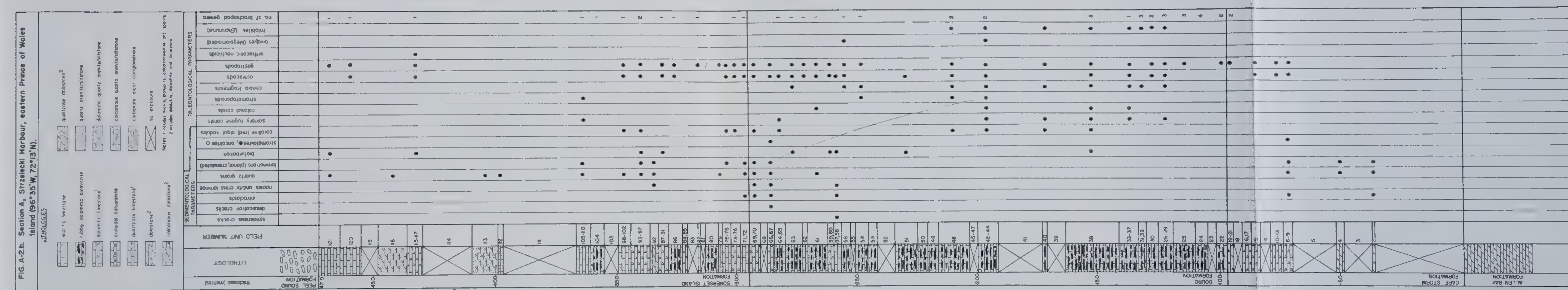


FIG. A-2b. Section A, Strzelecki Harbour, eastern Prince of Wales Island (96°35'W, 72°13'N).

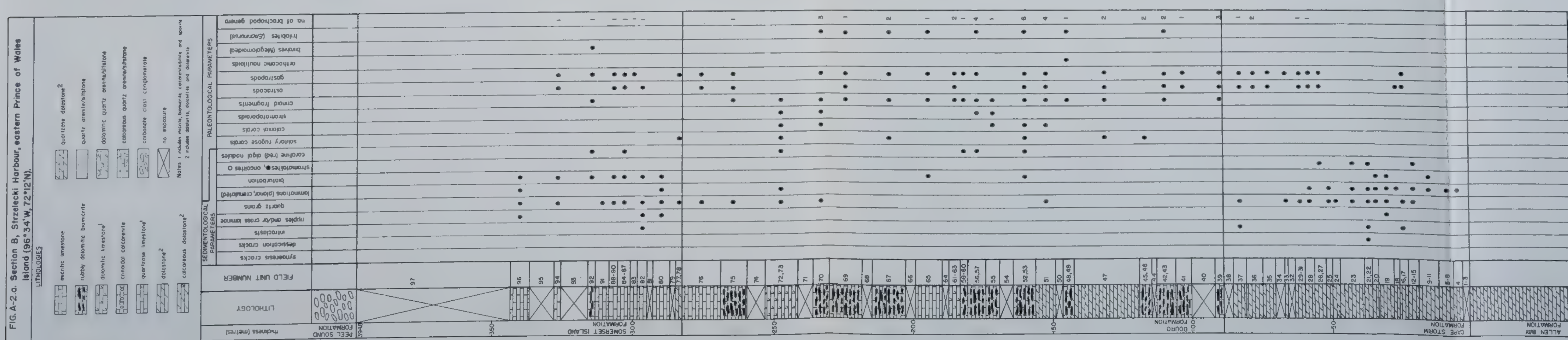


FIG. A-2a. Section B, Strzelecki Harbour, eastern Prince of Wales Island (96°34'W, 72°12'N).

FIG.A-3.a. Section O, Cape Brodie, eastern Prince of Wales Island (96°34'W, 72°23'N).

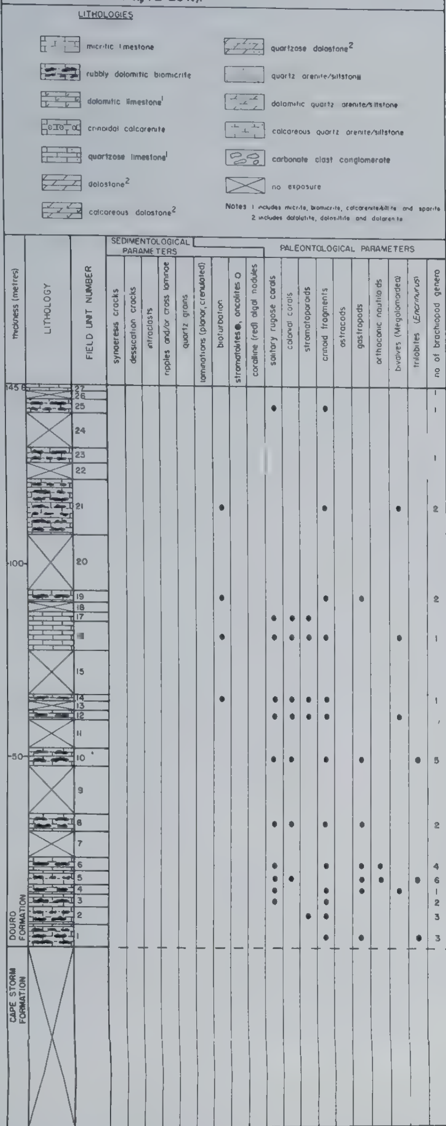


FIG.A-3.b. Composite section P and Q, "Flexure Bay", north Cape Brodie, eastern Prince of Wales Island (96°26'W, 72°33'N).

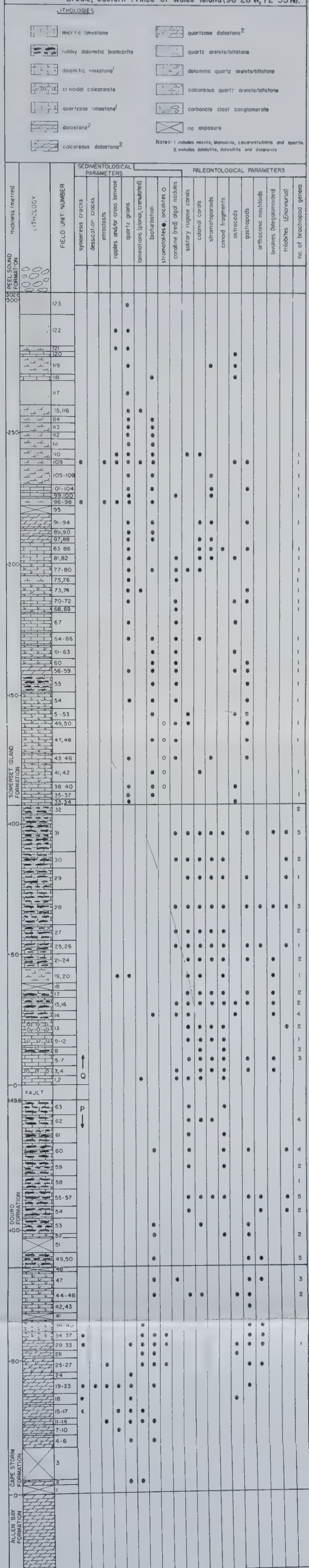


FIG.A-3.c. Composite section H and I, Savage Point, eastern Prince of Wales Island (96°34'W, 72°41'N).

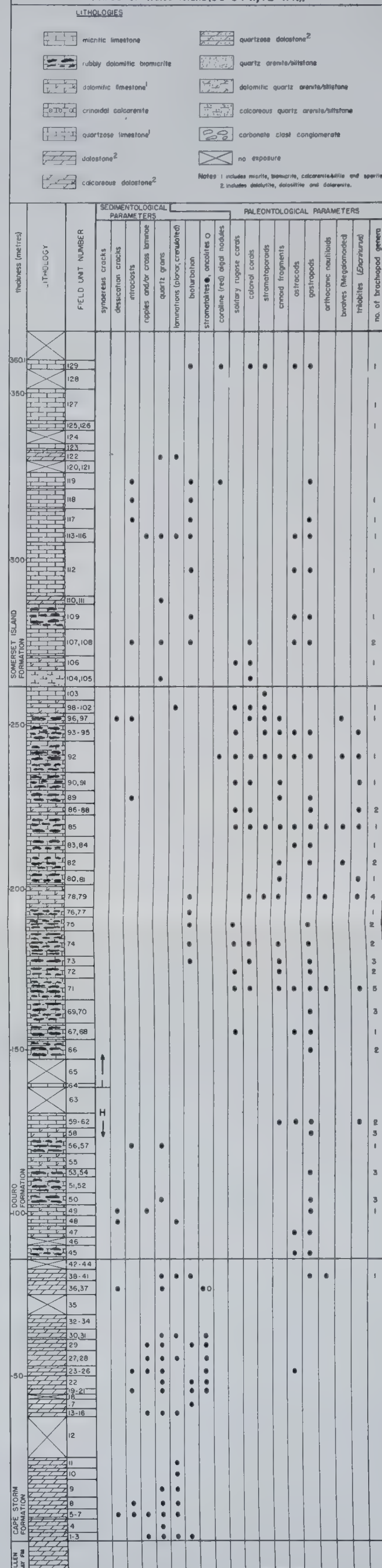


FIG. A-4.a. Section J, Savage Point, eastern Prince of Wales Island (96°41'W, 72°43'N).

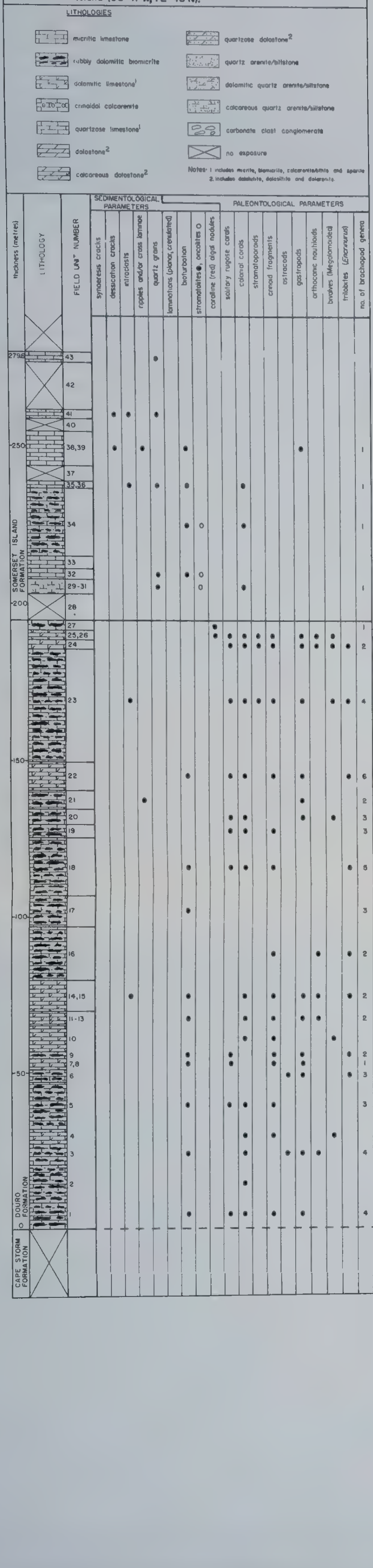


FIG. A-4.b. Section R, southern Prescott Island (96°46'W, 73°01'N).

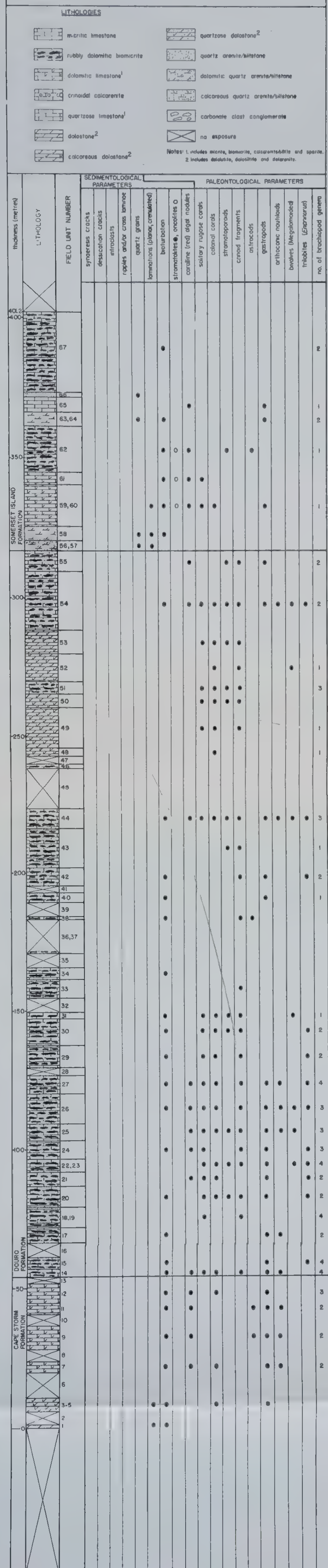
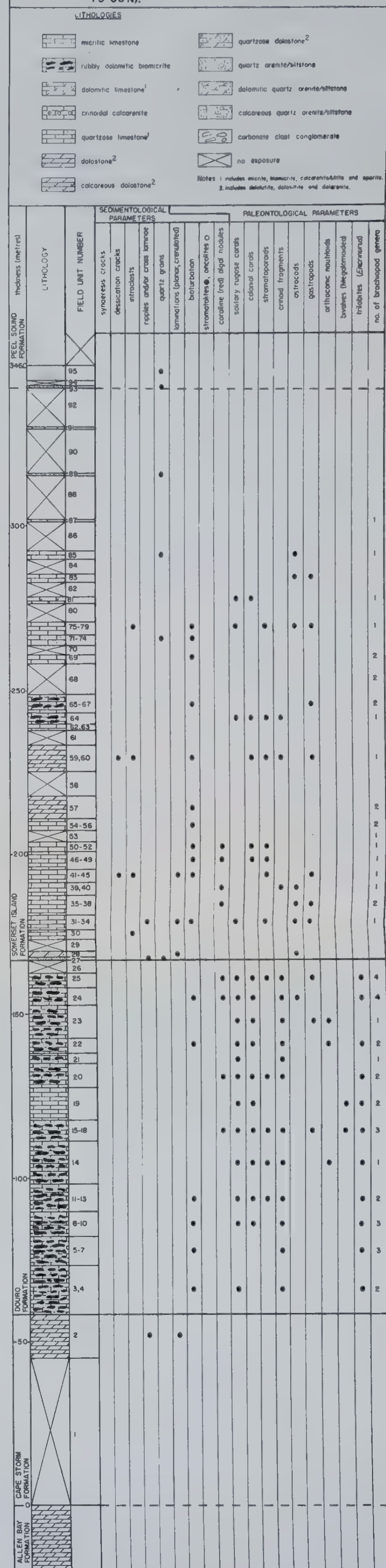


FIG. A-4.c. Section K, northern Prescott Island (96°49'W, 73°06'N).



XII. APPENDIX II, CONODONT IDENTIFICATION

Report No. 1-TTU-1984 Geological Survey of Canada

Report of 18 lots of conodont samples from the Douro and Somerset Island Formations on eastern Prince of Wales Island and Prescott Island (NTS 68A, D), District of Franklin; submitted by Mr. Paul Mortensen.

Field unit A-38, Douro Formation, 62.9-64.3 m above the base of the formation,
Strzelecki Harbour, Prince of Wales Island,
73°13''N, 96°13''W.

Conodonts
Oulodus sp. 3 of Uyeno (1981)
O. sp.
Ozarkodina confluens (Branson and Mehl) alpha morph
O. excavata excavata (Branson and Mehl)
Panderodus sp.

Geological Survey Location No. C-120875.

Field unit A-61, Douro Formation, 164.6-172.8 m above the base of the formation,
Strzelecki Harbour, Prince of Wales Island,
72°13'N, 96°13'W.

Conodonts
Ozarkodina n. sp. H of Uyeno (1981)
Pelekysgnathus arcticus Uyeno

Geological Survey Location No. C-120876.

Field unit B-43, Douro Formation, 23.3-24.0 m above the base of the formation,
Strzelecki Harbour, Prince of Wales Island,
72°12'N, 96°34'W.

Conodonts
Oulodus sp.
Ozarkodina confluens (Branson & Mehl) alpha morph
O. excavata excavata (Branson & Mehl)
Panderodus sp.

Geological Survey Location No. C-120877.

Field unit B-60, Douro Formation, 93.1-95.2 m above the base

of the formation,
Strzelecki Harbour, Prince of Wales Island,
72°12'N, 96°34'W.

Conodonts
Ozarkodina excavata excavata (Branson & Mehl)
Panderodus sp.

Geological Survey Location No. C-120878.

Field unit B-75, Douro Formation, 171.9-180.2 m above the base of the formation,
Strzelecki Harbour, Prince of Wales Island,
72°12'N, 96°43'W.

Conodonts
Ozarkodina n. sp. H of Uyeno (1981)
Pelekysgnathus arcticus Uyeno

Geological Survey Location No. C-120879.

Field unit C-55, Douro Formation, 70.6-72.1 m above the base of the formation,
Strzelecki Harbour, Prince of Wales Island
72°12'N, 96°33'W.

Conodonts
?Ozarkodina douroensis Uyeno (fragmentary Pa)
O. excavata excavata (Branson & Mehl)
Panderodus sp.

Geological Survey Location No. C-120880.

Field unit G-71, Douro Formation, 112.7-121.4 m above the base of the formation,
Le Feuvre Inlet, Prince of Wales Island,
72°16'N, 96°41'W.

Conodonts
4 acid-etched fragments, possibly of condonts.

Geological Survey Location No. C-120881.

Field unit G-78, Douro Formation, 178.6-187.6 m above the base of the formation,
Le Feurve Inlet, Prince of Wales Island,
72°16'N, 96°41'W.

Conodonts
Ozarkodina sp. indet. (2 badly etched fragmented Pa)
Panderodus sp.

Geological Survey Location No. C-120882.

Field unit G-82, Douro Formation, 195.6-203.0 m above the base of the formation,
Le Feurve Inlet, Prince of Wales Island,
72°16'N, 96°41'W.

Conodonts
Ozarkodina sp. (small Pa)
Pelekysgnathus arcticus Uyeno
Panderodus sp.

Geological Survey Location No. C-120883.

Field unit G-96, Somerset Island Formation, 15.0-18.6 m above the base of the formation,
Le Feurve Inlet, Prince of Wales Island,
72°16'N, 96°41'W.

Conodonts
Pelekysgnathus arcticus Uyeno

Geological Survey Location No. C-120892.

Field unit HI-64, Douro Formation, 52.2-53.0 m above the base of the formation,
Savage Point, Prince of Wales Island,
72°41'N, 96°34'W.

Conodonts
Oulodus sp.
Ozarkodina excavata excavata Branson & Mehl
Panderodus sp.

Geological Survey Location No. C-120884.

Field unit HI-84, Douro Formation, 128.0-129.5 m above the base of the formation,
Savage Point, Prince of Wales Island,
72°41'N, 96°34'W.

Conodonts
Oulodus sp.
Panderodus sp.

Geological Survey Location No. C-120885.

Field unit J-10, Douro Formation, 56.9-65.2 m above the base of the formation,
Savage Point, Prince of Wales Island,
72°43'N, 96°41'W.

Conodonts
Oulodus sp.
Ozarkodina confluens (Branson & Mehl)
O. excavata excavata (Branson and Mehl)

Panderodus sp.

Geological Survey Location No. C-120886.

Field unit J-16, Douro Formation, 79.6-96.7 m above the base
of the formation,
Savage Point, Prince of Wales Island,
72°43'N, 96°41'W.

Conodonts
Oulodus sp.
Ozarkodina confluens (Branson and Mehl)
O. excavata excavata (Branson and Mehl)
O. sp. indet. (large Pa)

Geological Survey Location No. C-120887.

Field unit K-11, Douro Formation, 32.4-36.2 m above the base
of the formation,
Northern Prescott Island
73°06'N, 96°49"W

Conodonts
Ozarkodina confluens (Branson and Mehl)
gamma morph transitional to epsilon morph
Panderodus sp.

Geological Survey Location No. C-120888.

Field unit K-24, Douro Formation, 88.7-93.9 m above the base
of the formation,
Northern Prescott Island,
73°06'N, 96°49'W

Conodonts
Apparatus B of Uyeno (1981) (Pb)
Ozarkodina confluens (Branson and Mehl)
alpha morph, and gamma morph
transitional to epsilon morph
O. excavata excavata (Branson and Mehl)
Panderodus sp.

Geological Survey Location No. C-120889.

Field unit L-48, Douro Formation, 0.0-2.2 m above the base
of the formation,
Strzelecki Harbour, Prince of Wales Island,
72°11'N, 96°34'W

Conodonts
Apparatus B of Uyeno (1981)
Oulodus sp.
Ozarkodina confluens (Branson and Mehl)
gamma morph, and gamma morph

transitional to epsilon morph
O. douroensis Uyeno
Panderodus sp.

Geological Survey Location No. C-120890.

Field unit L-70, Douro Formation, 73.5-75.4 m above the base of the formation,
Strzelecki Harbour, Prince of Wales Island,
72°11'N, 96°34'W

Conodonts
Oulodus sp.
Ozarkodina confluens (Branson and Mehl)
gamma morph
O. douroensis Uyeno
Panderodus sp.

Geological Survey Location No. C-120891.

Based on the work of Thorsteinsson and Uyeno (1981), the majority of the conodonts listed above are consistant with the assignment of the enclosing rocks to the *siluricus* Zone of late Ludlovian age. There are two exceptions: GSC locs. C-120876 and C-120879 (Field nos. A-61 and B-75) which contain *Ozarkodina* n. sp. H of Uyeno (1981), and which occur high in the Douro Fm., only a short interval below the overlying Somerset Island Fm. (=lower member, Peel Sound Fm. of Thorsteinsson, 1981). *O.* n. sp. H was previously recovered from the Somerset Island Fm., (Somerset Island), Barlow Inlet Fm. [Cornwallis Island, and there above the first appearance of *Pedavis latialata* (Walliser)], and the lower member of the Peel Sound Fm. (eastern Prince of Wales

Island), all in the *latialata* Zone.

Since *Ozarkodina* n. sp. H is apparently confined to the *latialata* Zone over a relatively wide area, its occurrence in the higher parts of the Douro Fm. may perhaps suggest that the top of the formation may have been picked at different levels (and hence on different criteria) by different workers. If its occurrence in the Douro Fm. is accepted, however, the only logical conclusion is that (1) the range of *O.* n. sp. H extends down into the upper part of the *siluricus* Zone (*Osarkodina douroensis*, a *siluricus* Zone indicator, occurs in the uppermost 1.82 m of the Douro Formation on Somerset Island; see Thorsteinsson and Uyeno, 1981, Fig. 18, p. 25); or (2) the top of the Douro Fm. is diachronous.

All morphotypes of *Ozarkodina confluens* listed above are those introduced by Klapper and Murphy (1975).

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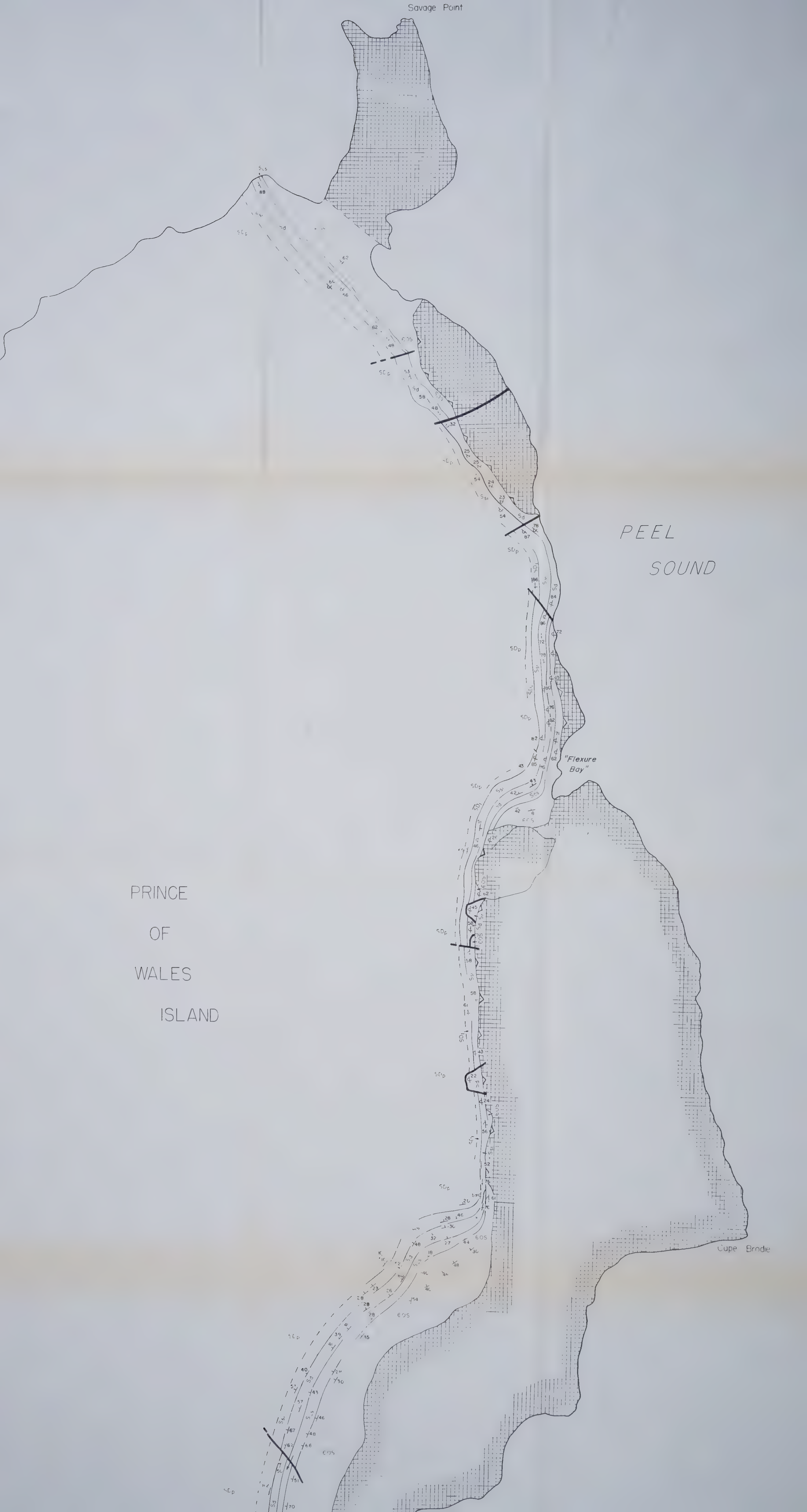
FIG. 2.1.

Geology of cent

Prince of Wales

P.


Savage Point
PEEL
SOUND
"Flexure
Bay"
PRINCE
OF
WALES
ISLAND
Cape Brodie

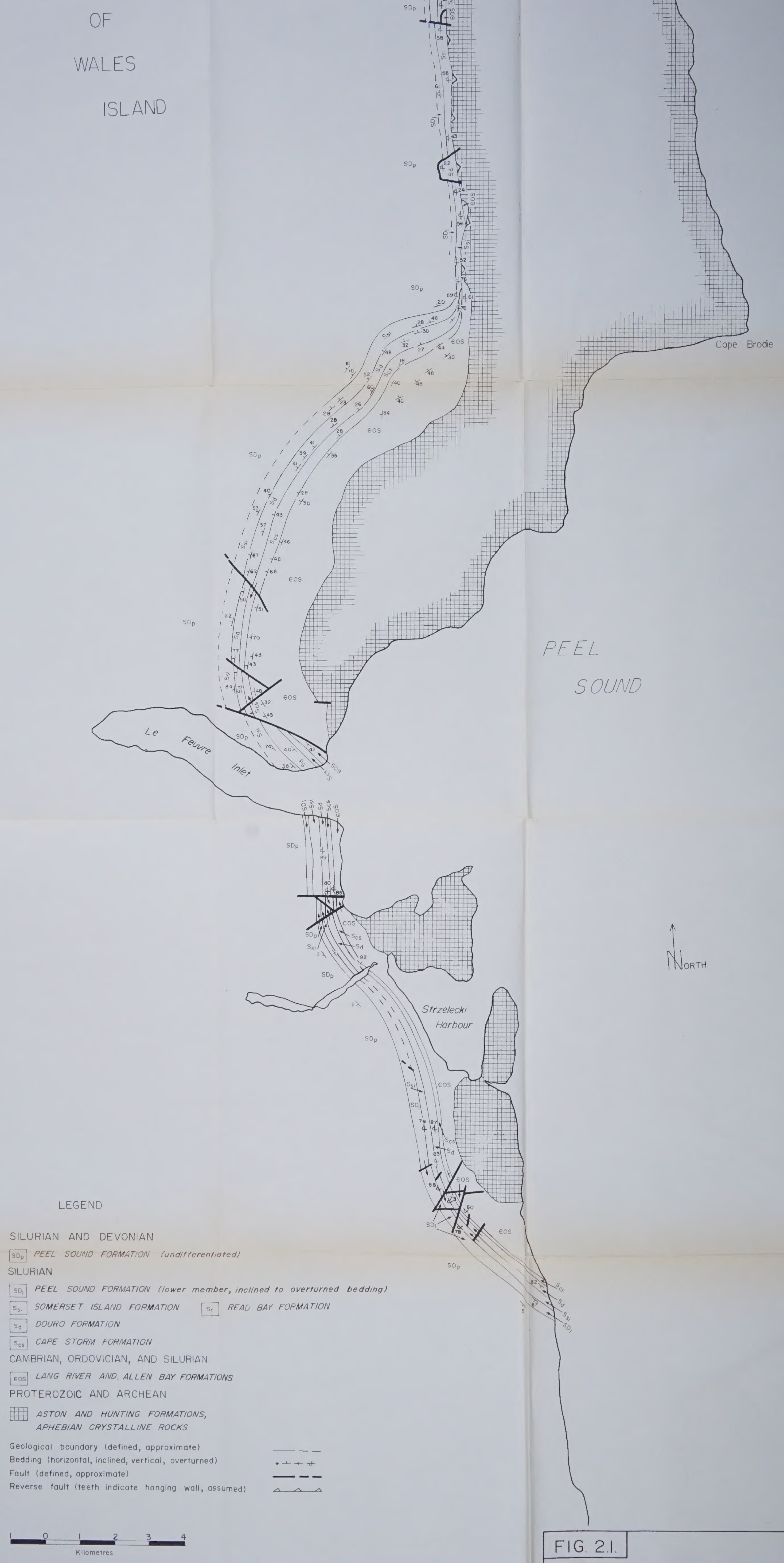


FIG. 2.1. Geology of central-eastern Prince of Wales Island.

P. Mortensen, Oct., 1983


VIVIAN
ISLAND
PEEL
SOUND
PRESCOTT
ISLAND
PEEL
SOUND
Cape M'Clure
LEGEND
SILURIAN AND DEVONIAN
SDp PEEL SOUND FORMATION (undifferentiated)
SILURIAN
Ssi SOMERSET ISLAND FORMATION
Sr READ BAY FORMATION
Sd DOURO FORMATION
Scs CAPE STORM FORMATION
CAMBRIAN, ORDOVICIAN, AND SILURIAN
€OS LANG RIVER AND ALLEN BAY FORMATIONS
PROTEROZOIC AND ARCHEAN
ASTON AND HUNTING FORMATIONS, APHEBIAN CRYSTALLINE ROCKS
Geological boundary (defined, approximate)

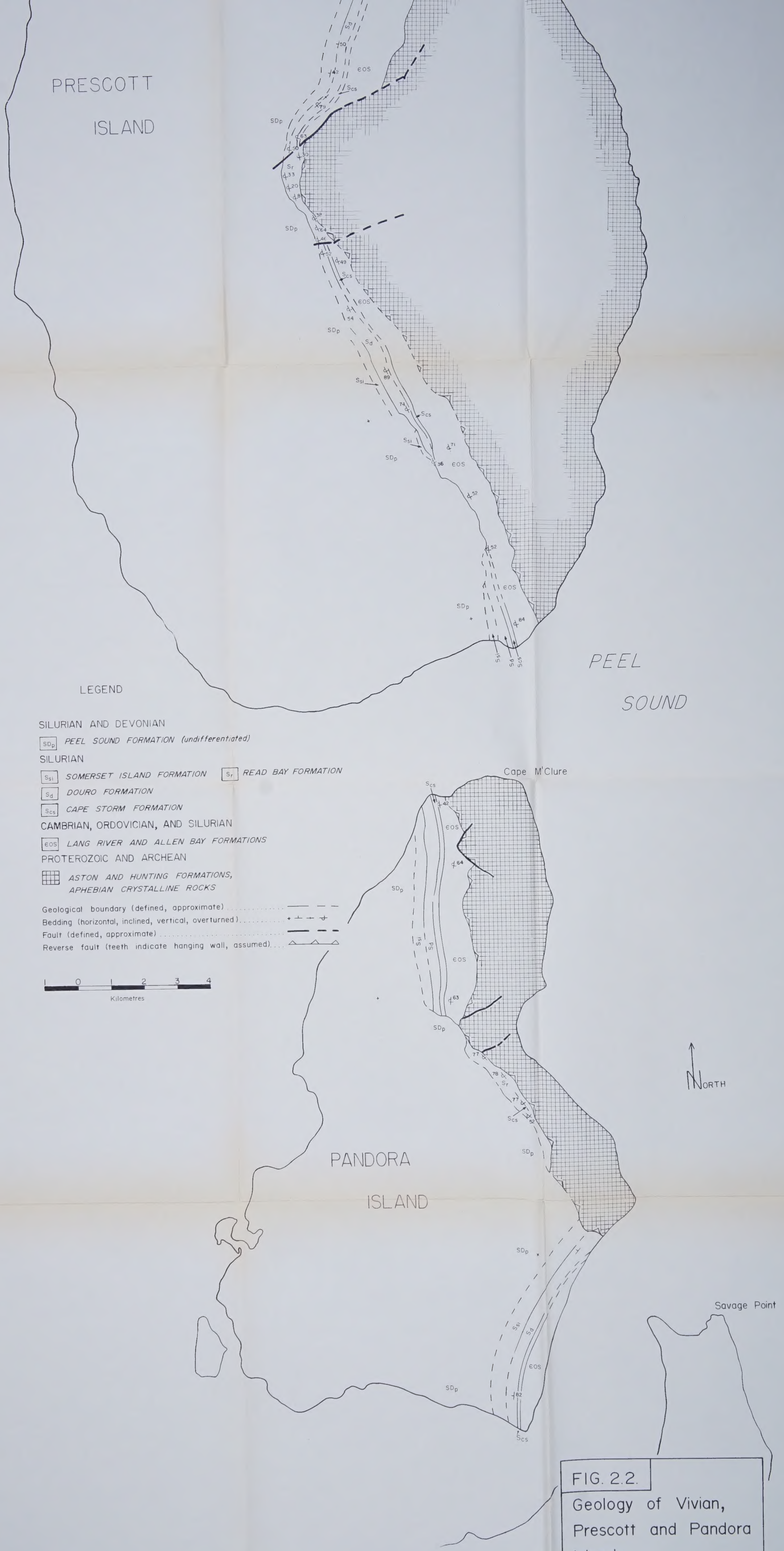


FIG. 2.2. Geology of Vivian, Prescott and Pandora Islands.

P. Mortensen, Oct., 1983